Mid to late Holocene sea level history and coastal evolution in Tasmania

by

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Abstract

An understanding of past sea level change in Tasmania is essential to understanding the contribution of Antarctic ice melt to the global oceans. Research into Holocene sea level history in Tasmania has been sporadic and conflicting. The primary goal of this research was to develop precise Holocene sea level index points (SLIPs) for Tasmania, south eastern Australia, achieved by: establishing the relationship between Tasmanian coastal salt marsh biogeomorphology and mean sea level; and locating and documenting palaeo-environments that contain a Holocene sea level signal of substantial duration. The northern hemisphere quantitative, base of basal salt marsh approach was combined with the Australian coastal geomorphic landform evolution approach to reconstruct Holocene sea level history. Pleistocene transgressive surfaces in micro tidal drowned river valleys in south eastern Tasmania provided a set of depth sequenced base of basal sea level index points. A full intertidal foraminifera based transfer function was calibrated with a sediment core to fill gaps in this sea level record. A contrasting coastal evolution sedimentary record in a meso tidal, open coast intertidal setting in north western Tasmania was examined by targeting an organically enriched sedimentary deposit underlying the modern salt marsh, that was inferred to be a drowned Holocene salt marsh facies.

Tasmanian salt marshes are constrained to elevations relative to MSL within sites, but inundation period is a better predictor of marsh zonation than elevation. The greatest between site variation in inundation period occurs at marsh seaward edges that are limited to no more than 33% tidal inundation period per year. This regional variation in inundation period of the seaward edge limits the suitability of regional modern analogues for transfer function based sea level reconstructions.

Sea levels were below present throughout the Holocene, with a mid-Holocene highstand absent. The slow rates of sea level rise after 6000 yr BP indicate continued melt water contribution from Antarctica after 6000 yr BP to the present. The open coast site in north western Tasmania did not yield a longer Holocene sea level history. The age of the putative marsh was from ~ 37 000 yr BP to ~ 26 000 yr BP. It was a preserved, semi indurated Pleistocene swamp deposit that

occupied the swales between sand dune ridges through the latter stages of glacial advance, and just prior to the last glacial maximum, when aeolian processes dominated the landscape.

Integrating the sea level reconstruction for Tasmania with the accepted geochronological framework for estuarine evolution added further detail on the geomorphic evolution of coastal landforms in south eastern Australia, particularly the prevalence of the transgressive sand sheet, now shown to be present in shallow marine environments outside of estuaries. Further research to better populate the early and late Holocene sea level curve for Tasmania should focus on locating the transgressive surface in shallow subtidal, and upper intertidal locations.

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Chapter 1 Introduction

1.1 Context

Sea level research in Australia has been temporally and spatially sporadic in comparison with other regions such as the U.S Atlantic Coast (Horton et al., 2007; Engelhart et al., 2009, 2011b; Horton et al., 2009; Engelhart and Horton, 2012; Kemp et al., 2014) and the British Isles (Horton and Edwards, 2006; Shennan and Horton, 2002; Shennan et al., 2006 & 2009; Horton and Shennan, 2009; Shennan et al., 2012), where relative sea level change is now well understood. In these regions, a wealth of high resolution relative sea level change studies now inform regional glacial – isostatic adjustment models which collectively isolate the eustatic sea level signal (Engelhart et al., 2011b; Kemp et al., 2014; Peltier et al., 2014; Kopp et al., 2016). Satellite altimetric observations of change in the volume of the global oceans also inform understanding of sea level change (Engelhart et al., 2009; van der Wal et al., 2015; Watson et al., 2015).

The southern hemisphere lacks the detail of sea-level reconstruction that gives the longer term Holocene background sea level changes to support the more recent satellite altimetry and tide gauge data, and recent sea level change reconstructions (Gehrels and Woodworth, 2013). This information is critical to accurately identify future risks to coastal landforms, populations and infrastructure with any sea level change (Gehrels and Long, 2008; Siddall and Milne, 2012; Kemp et al., 2014; Lambeck et al., 2014; Watson et al., 2015).

Few Australian studies have been specifically designed to develop sea level index points from finite intertidal indicators. Thus, sea level history is ambiguous (Murray-Wallace and Woodroffe, 2014). Development of a regional glacio - hydro - isostatic adjustment (GHIA) model has not progressed since the first iteration by Lambeck and Nakada in 1990, despite three major critical reviews of sea level change and coastal evolution studies since this time (Murray-Wallace 2002; Sloss et al., 2007; Lewis et al., 2013). Both reviews indicated that more accurate sea level index points (SLIPs) were required. This increase in accuracy would allow improved understanding of glacial and hydro – isostatic adjustment for the region (Engelhart et al., 2009; Lambeck, 2002; Lambeck et al.,2010; Lambeck et al., 2014), and is critical in calculating the contribution of glacial meltwater to the volume of the global oceans, where debate still exists as to the timing, magnitude and relative contribution of the major polar, and other land based ice sheets, (Nakada and Lambeck, 1989; Lambeck and Nakada, 1990; Shennan and Horton, 2002; Peltier et al., 2002; Milne et al., 2005; Gehrels and Long, 2008; Long et al., 2012; Lambeck et al., 2014).

To help address this significant gap in the Australian and global data set, the present work establishes a mid to late Holocene sea level curve for Tasmania, Australia. As a southern hemisphere far field location, remote from former ice sheets, Tasmania provides an important location for the eustatic sea level signal (Nakada and Lambeck, 1989; Lambeck et al.,, 2010). It has a narrow continental shelf so not subject to continental levering as the proximal ocean basin fills, and is tectonically stable at least for the recent past (Murray-Wallace and Goede, 1995). Furthermore, being at the southern end of the Australian continent, the sea level history is a critical input to the regional glacio –hydro isostatic adjustment model. However, little is known of Tasmanian sea level history. The sole accurate reconstruction only provides evidence for the past 650 years before present (Gehrels et al., 2012). Other previous work provides limited stratigraphic evidence of sea level change and concomitant coastal evolution. Limited biogeomorphic evaluation of the modern environments has taken place to aid understanding of palaeo-environments (Clark et al., 2011; Gehrels et al., 2012).

Most existing relative sea level research has been carried out where considerable pre-existing knowledge informs the sea level reconstruction approach (Shennan and Horton 2002, van de Plassche, 1986; Engelhart et al., 2009 & 2011), including of the biological response of indicator organisms to the physical environment. The most precise indicators are those located in the sedimentary depositional environments of saltmarshes (Van de Plassche, 1986; Gehrels, 2002; Horton and Edwards, 2006; Engelhart et al., 2009 & 2011a). These provide a range of lithological and ecological indicators that preserve as an archive of past environments represented by their micro and macro fossils. However, their modern ecology should be fully known to interpret their response to sea level and other environmental change.

In the northern hemisphere well over a century of ecological and geomorphological research in salt marshes has informed the development of palaeo sea level reconstruction (Birks and Birks 1980; Berglund, 2003; Haslett, 2001; Engelhart et al., 2009; Birks et al., 2014). Tasmania has approximately 39 km^2 of native salt marsh around its coastline in sheltered bays and drowned river valley estuaries (Zann, 1996). However, the bio-geomorphic relationship to mean sea level is not known, nor their regional specific evolutionary pathway, and these features needs assessment, before appropriately targeted methods of reconstruction can be determined (Gehrels, 2002; Horton and Edwards, 2006; Wright et al., 2011).

On the east coast of Australia observational evidence of relative sea level change is consistent with the GHIA model (Lambeck and Nakada 1990). An observed mid Holocene highstand diminishes in magnitude from north to south along the east coast of the Australian continent (Lewis et al., 2013). For Tasmania the GHIA model predicts that sea level was at, or slightly above its present level, of around 25 cm at its Holocene maximum (Lambeck and Nakada, 1990; Lambeck, pers.comm. in Gehrels et al., 2012), but its existence is not clear from previous work, so this quantum needs verification from sea level index points (SLIPs) in Tasmania (Lambeck and Nakada 1989; Lambeck at al., 2010; Gehrels and Woodworth, 2013).

Due to the small expected sea level change over the mid to late Holocene, the local SLIPs need to be of high precision with error ranges within the likely variation in sea levels over the period. In micro tidal dominated Tasmania, such index points are likely to be located within quiescent estuarine zones, because there the processes of deposition recorded in the stratigraphic sequence should capture the sea level signal by tidal, rather than wave dominated processes. The challenge of reconstructing past sea levels for the mid to late Holocene in Tasmania is likely to require a variety of methods, to piece together evidence of environmental change caused by physical factors linked to mean sea level (Berglund, 1986; Starkel, 1998).

1.1 Thesis Aim and structure

The primary aim of this thesis is to establish the nature of mid to late Holocene sea level history for Tasmania. The two primary objectives are to create precise south eastern Australian SLIPs, for input to a new glacio isostatic adjustment model, and to provide a background Holocene sea level record that will inform understanding of recent rates of sea level change (Gehrels and Woodworth 2013; Lambeck et al., 2014). In order to address the primary aim, the thesis commences with a literature review and is followed by four chapters written in academic journal format.

Chapter 2 The causes of Holocene sea level change and how it has been detected.

The literature review examines the causes of global, regional and local relative sea level change. It examines the current knowledge of Australian sea level history from the observational record, how it relates to the region's glacio – isostatic adjustment model (Lambeck and Nakada, 1990), and what it means for Tasmanian sea level history. Because some authors suggest Tasmania experienced higher than present Holocene sea levels, and some observed no such evidence, the review describes the types of sedimentary sequences that might be encountered in the sedimentary record in Tasmania, including transgressive and regressive, and highstand features. Finally, what is known of salt marsh evolution pertinent to sea level studies are examined, including how these might vary locally or regionally, because little is currently known of Tasmanian salt marshes and their relationship to mean sea level, from which Holocene sea level research can commence.

Chapter 3 Regional variation in biogeomorphological zonation of Tasmanian salt marshes

The aim of Chapter 3 is to quantify the modern relationship between salt marsh biogeomorphology, mean sea level and measurable abiotic variables, in three different geomorphic settings and climates and two tidal ranges, from north western to south eastern Tasmania. The regional approach is taken to identify the causes of variation to elevation zonation and achieves its objective of informing the later palaeo sea level reconstruction work by linking biogeomorphic features to inundation period. The regional approach is also intended to make clearer the generalities around salt marsh form and process that are useful for general conceptual understanding, but of limited value for sea level studies from regions where little is known.

The study employs a statistical model for predicting the probability of marsh zone presence by tidal inundation period and relevant edaphic conditions across environmental ranges. Testing the prediction of marsh biogeomorphic features was considered pertinent for 3 reasons, 1) because quantitative palaeoecological research relies on prediction by quantitative estimates of past environments from proxies (Jordan, 2011), 2) because different statistical models should be explored as means of predicting species distribution from environmental variables to add perspective to existing knowledge of complex interrelation and correlation structures (Ahmadi-Nedushan et al., 2006), and 3), because it also provides a baseline of Tasmania salt marsh features which is an important contribution to the global salt marsh knowledge base that is currently limited to mainland Australian locations (Saintilan, 2009). The field data collection for Chapter 3 followed extensive reconnaissance of Tasmania's salt marshes and related depositional environments, and informed study sight selection for the following palaeo-reconstructions comprising Chapters 4, 5 and 6.

Chapter 4 A mid Holocene sea level record for the Little Swanport Estuary, south eastern Tasmania

The aim of Chapter 4 is to improve the accuracy of the mid Holocene sea level history for Tasmania and examine whether it conforms to other records, and the GHIA model prediction. Previous studies are examined that give proxy sea level information, inferred from coastal landform evolution, with few direct sea level index points, leading to debate over the presence or absence of a mid-

Holocene sea level highstand. The key research question is; was there a mid-Holocene sea level highstand in Tasmania?

The study presents 10 new sea level index points retrieved from the contact between the transgressive surface of a shallow low relief drowned river valley and the overlying estuarine transgressive sand sheet. It evaluates SLIPs from two other studies conducted in the region (Clark et al., 2011; Gehrels et al., 2012), and incorporates them into the data set to create a sea level curve for the late stages of the early Holocene, to the end of the mid Holocene. It uses traditional qualitative methods of sediment stratigraphy that is described within a geochronological framework for coastal evolution studies in south eastern Australia, developed by Sloss et al., (2004; 2006a, b; 2010), to extend the knowledge base of intertidal estuarine evolution for the region.

Chapter 5 Late Holocene sea level stability and very slow sea level rise in Tasmania, south eastern Australia: A full intertidal foraminifera transfer function broadly indicates the nature of Late Holocene sea level.

The aim of Chapter 5 is to identify the nature of late Holocene sea levels in Tasmania, and examine whether it follows the mid Holocene sea level trend identified in Chapter 4. The key research question generated from Chapter 4 is whether sea levels continued to rise very slowly, or were at a still stand. This is important because Tasmanian sea levels are expected to most closely follow the eustatic sea level (Nakada and Lambeck, 1989; Lambeck and Chappell 2010). The contribution is most important to provide SLIPs for the GHIA, but also to understanding the nature of coastal evolution in Tasmania.

To achieve this aim a full intertidal foraminifera based transfer function is tested to extend the existing 650 yr BP salt marsh record, as far back to the late Holocene as can be achieved. The primary objective is to evaluate the suitability of a full intertidal transfer function, to develop late Holocene SLIPs for Tasmania. The transfer function is calibrated with a sediment core to broadly indicate the nature of sea level change and identify appropriate sampling locations for further chronological work.

Chapter 6 Late Pleistocene and Holocene intertidal stratigraphy of Boullanger Bay north western Tasmania: Seeking an early mid to Holocene sea level record.

The last substantive chapter seeks to extend the Holocene sea level record by examining stratigraphic evidence from an open coast, meso tidal setting. Having the largest tidal range on the island, the site presented an extensive intertidal area with modern salt marsh overlying a deep facies that was inferred to be a drowned salt marsh deposit. The aim of the study is to constrain the age of the deeper deposit and understand its geomorphic evolution and potential to provide a precise early to mid-Holocene sea level record. The stratigraphy is interpreted within the geochronological framework for coastal evolution in south eastern Australia (Sloss et al., 2004; 2006a, b; 2010), employed in Chapter 4, and is correlated with other palaeo-environmental studies in the region to understand the landform's Holocene geomorphic evolution.

Chapter 7 Conclusion

Chapter six briefly draws together the key findings of the thesis, discusses the limitations to the study, and makes recommendations for future work.

Chapter 2

Literature Review: What cause changes to relative sea level and how is it recorded?

2.1 Introduction

Sea level change is variable in time and space (Lambeck and Chappell, 2001; Whitehouse, 2009; Church et al., 2010; Engelhart et al., 2011; Murray-Wallace and Woodroffe, 2014; Shennan et al., 2015) so sea level studies are conducted at various scales. Sea level changes of hundreds of meters over millions of years have been generalised to patterns of monotonic rise and fall across continents. At this scale, the relevant driving processes are plate tectonics, mountain building, and sea floor spreading (Lambeck and Chappell, 2001; Zachos et al., 2001; Coe, 2003; Milne et al., 2006; Lambeck et al., 2014). At regional scales, sea level research focusses on time scales of 10^{-3} to 10^{-5} years, with the relevant forcing factors dominated by glacio - hydro – isostatic adjustment of the Earth surface to ice sheet expansion and contraction through Croll-Milankovitch climate cycles (Lambeck and Chappell 2001, Yokayama et al., 2001; Zachos et al., 2001; Peltier, 2002; Coe, 2003; Siddal and Milne, 2003; Shennan, 2006; Engelhart et al., 2011; Catuneanu, 2011). Superimposed on such global climate change scales and concomitant greater planetary deformation are local scale boundary conditions that effect relative sea level changes including, antecedent topography, size of the receiving basin, sediment supply and type, meteorological forcing, and neo-tectonics all of which have been observable in sedimentary records of the Holocene Epoch (Bird, 2000: Lambeck and Chappell, 2001; Zachos, 2001; Woodroffe, 2002; Church et al., 2010).

Undertaking sea level change research requires an understanding of 1) the causes of sea level change, 2) how to find a sea level change record, and 3) how the record contributes to the local, regional and global sea level history data base. This review identifies the causes of sea level change within temporal ranges including the Quaternary Pleistocene climates, Holocene and anthropogenically induced climate changes, because all of these are likely to be identified in the sedimentary record in Tasmania, where a sea level curve is absent. To understand how to find a sea level change record the review focusses on the types of Holocene sea level change sequences evident in Australia, focussing on estuarine sediment facies, where the Holocene record is most likely to be found in Tasmania. Finally, the review examines how salt marshes are used to develop sea level records, describing the global scale range of biogeomorphic features pertinent to sea level research to compare with Tasmanian marshes, for which little is known.

The Holocene Epoch commenced 11,700 yr BP (Walker et al., 2012) following the Last Glacial Maximum at 21,000 yr BP, when sea levels were approximately 120 m lower than present. Commencement of deglaciation occurred around 18,000 years BP, when approximately 50 million km^3 of land-based ice began to melt (Peltier and Fairbanks, 2006). Sea levels rose rapidly from this time and began slowing around 8,000 yr BP. The early Holocene is considered to be between 11.700 yr BP and 8,200 yr BP (Walker et al., 2012). Glacial meltwater addition to the oceans all but ceased at approximately 6,000 yr BP when sea levels rose to approximately present levels. During the mid Holocene (8,200-4,200 yr BP) rheological adjustments to the redistribution of water through from the cryosphere to the hydrosphere occurred and typified by periods of higher levels than present from approximately 7000 – 5000 yr BP, because flexure of the sea floor continued after melt water mostly ceased, (Lambeck, 2002; Walker et al., 2012; Lambeck et al., 2014). The late Holocene commencement is marked by sea level fall from a high-stand in many places (Walker et al., 2012), and this sea level record is dominated by the time dependant readjustment of the stable water / ice load ratio (Lambeck, 2002), which is still active today (Engelhart et al. 2011).

Sea level change is measured as a change in the elevation of mean sea level, relative to the Earth’s surface (relative sea level). Relative sea level can fall when the height of the ocean falls or the land surface rises. The height of the ocean may rise when the shape of the geoid changes, due to there being an increase in the volume of water in the ocean, or a decrease in the storage capacity of the ocean. Eustatic sea level change is described as the ice equivalent term for the addition of melt water from diminishing glaciers (Lambeck, 2002). Other factors contributing to eustatic sea levels include the steric contribution (change in ocean water density due to change in temperature and salinity) and the geoid contribution of gravitational and rotational changes due to the mass balance of ocean water to ice (Engelhart et al., 2011).

Changes in the height of the surface of the Earth occur also with tectonic displacement or from sediment, water or ice loading (Murray-Wallace and Woodroffe, 2014). All change in sea level excepting that caused by tectonics is driven in some way by climate because of its influence on the relative volume of ice / water and the rate at which sediment is eroded, transported and deposited in landform building and maintenance processes.

Interpreting relative sea level history requires understanding the forcing factors controlling the volume of the global oceans and the rheological adjustments of the Earth's crust to the change in the distribution of water, termed (isostasy).

2.2 Causes of sea level change

2.2.1 Quaternary Pleistocene climates with orbital forcing

Climate change has been the key forcing factor of sea level change during the Quaternary Period, affecting the mass balance of glaciers and ice sheets with the global oceans (Church et al., 2010; Kemp et al., 2014; Lambeck et al., 2014; Peltier, 2015). The Croll-Milankovitch Hypothesis explains changes in the geometry of Earth's orbit around the sun which controls the intensity of insolation received at Earth's outer atmosphere (Bradley, 1999). This 'astronomical theory' of ice ages describes three variable components of orbital geometry that are relevant to Quaternary climates and changes in sea levels.

These include eccentricity, obliquity and precession (Hays et al., 1976; Williams et al., 1998; Bradley, 1999). There has been continuous improvement in understanding of the internal climatic feedbacks and lag anomalies of the ocean, air, and ice system, building on the Croll-Milankovitch Hypothesis (Hays et al.,1976; Ruddiman and McIntyre, 1981; Ruddiman et al., 1986; Huybers and Wunsch,2005; Ruddiman, 2006; Abe-Ouchi et al., 2013)

Eccentricity refers to variation in the short axis of the Earth's slightly elliptical orbit around the sun (Williams et al., 1998). This results in the Earth being the closest to the sun (perihelion) around January 3rd which results in approximately 3.5% more solar radiation outside the atmosphere than the annual average. At June 3rd, when it is furthest (aphelion), it receives approximately 3.5% less than the annual average (Bradley, 1999). Geometric fluctuation of the elliptical orbit occurs quasi periodically, with an average period of 95,800 years over the past 5 million years. The eccentricity range within each period is from nearly circular at its minimum with almost no difference between perihelion and aphelion, to as much as 30% difference when eccentricity is at its greatest (Bradley, 1999).

Obliquity refers to the Earth's tilt on its rotational axis from the ecliptic plane on which it revolves around the sun. It is the primary control on the geographical distribution of insolation (Williams et al., 1998). It follows that fluctuating obliquity redistributes surface air temperature on a scale that impacts global atmospheric and ocean circulation, hence major climate changes ensue. The Earth's axis is currently at 23.44°, which varies from 21.8° to 24.4° over cycles of ~ 41,000 years (Murray-Wallace and Woodroffe, 2014). The most recent maximum angle of inclination occurred about 100 000 years ago (Bradley, 1999).

Precession is movement or "wobble" in the Earth's rotational axis. This results in occurrences such as the timing of the perihelion with respect to the seasonal fluctuation (Bradley, 1999) over approximately 20, 000 year periods (Williams, 1998; Huybers and Wunsch, 2005). While the “wobble” is independent of eccentricity, the effects of precession on the timing and radiation receipts at the equinoxes can be modulated by it. For example, at minimum eccentricity, the effect on the seasonal variation is almost negligible, however at its maximum, the 30% variation between summer and winter insolation has considerable impact on climatic variations (Bradley, 1999).

It is the variable interactions between the three orbital components that affect Earth’s climate (Williams, 1998; Bradley, 1999). The lower latitudes are most impacted by the fluctuations of eccentricity and precession, but the higher latitudes are most influenced by obliquity. This leads to asymmetry between hemispheres as eccentricity and precessional effects in each hemisphere are opposite, but obliquity effects are not (Bradley, 1999). These combined effects are complex in time and space, yet the timing and magnitude of the growth and retreat of continental ice sheets is broadly predictable.

Recent advances in understanding the relative influence of the three parameters account for the climate – Earth feedback system interaction with obliquity (Abe-Ouchi et al., 2013). The 100, 000 year climate cycle is driven by the geographic and climatological setting of the North American ice sheet and the variable insolation it receives, and modulation of the summer insolation variation in the precessional cycle with eccentricity. The last ice age ended around 12000 years ago. The Earth is currently in an interglacial period.

2.2.2 Holocene climate change

The Holocene Epoch is characterised by climatic amelioration and overall deglaciation (Murray-Wallace and Woodroffe, 2014) contrasting sharply with the late Pleistocene in terms of warmer mean temperatures and reduced variability (Birks et al., 2014). This climate stability coincides with eccentricity being less than at any other time over the last 5,000,000 years (Richardson et al., 2011). The factors affecting Holocene climate at 10^1 to 10^3 time scales are the same as for the Pleistocene, although the orbitally driven changes that induced the abrupt climate change from the late Pleistocene to the Holocene are relatively small and gradual (Birks et al., 2014).

Although orbital fluctuation caused little change in global insolation receipts, it did force significant seasonal and latitudinal redistribution of energy in both hemispheres (Luciene Dias de Melo and Marengo, 2008). Precessional fluctuation at the early Holocene led to perihelion at the Northern Hemisphere summer solstice, thus higher insolation at all Northern Hemisphere latitudes. As occurs with precessional geometry, the opposite occurred in the Southern Hemisphere but the impact was much smaller, leading to higher than previous global insolation there. Globally, insolation has continually decreased for the last 12000 years and Earth is now at the opposite of the precessional cycle, (Birks et al., 2014).

Understanding of the relationship between the global insolation budget and climate has further developed since the emergence of fully coupled ocean-atmosphere-biosphere models (Birks et al., 2014). These same models have been used to demonstrate anthropogenic contribution to natural levels of greenhouse gasses, which have been shown to influence the global insolation budget and climatic events.

2.2.3 Anthropogenically induced climate change

Anthropogenic climate change coincides with the period of industrialisation (Kutzbach et al., 2011) and primarily was caused by deforestation, other land use change (Luciene Dias de Melo and Marengo, 2008) and fossil fuel combustion (Boden et al., 2015). Carbon dioxide is vital in the regulation of the Earth’s surface temperature through radiative forcing and the greenhouse effect. Since the industrial revolution, an increase of 26% of carbon dioxide in the atmosphere has been observed (Bengtsson, 1994). Current CO_2 levels are over 400 ppm and are rising at approximately 2 ppm/year. The rate of change continues to accelerate (IPCC, 2014), with accelerated deforestation being a major contributor (Leahy, 2018).

Deforestation can be described as the mass clearing of forest vegetation. Deforestation results in high levels of carbon dioxide being released into the atmosphere and reduced levels of carbon dioxide being removed. Fossil fuel emissions are sourced from the burning of petroleum, coal, and natural gas and the production of cement. In 2009, liquid and solid fuels accounted for 76.6% of the emissions from fossil fuel burning and cement production (Boden et al., 2015). Figures from the Carbon Dioxide Information Analysis Center show cement production to have doubled since the 1990s and contributed 4.7% of global anthropogenic CO_2 in 2009 (Boden et al., 2015).

Increasing levels of CO_2 and other greenhouse gases (such as methane from intensive agricultural practices) cause increases to the Earth’s mean temperatures. While many consequences are apparent from increasing surface temperature, most relevant to this discussion are rising sea levels from the melting of glaciers and ice shelves, addition of that water to the oceans, and thermal expansion of ocean water (Church et al., 2013; Horton et al., 2015). Tide gauge data and proxy sea level records in combination have shown accelerations and inflections in modern rates that correlate with the warming climate (Gehrels and Long, 2008; Gehrels and Woodworth, 2013). The modern sea level rise rates are greater than mid to late Holocene rates which are referred to in this context as background rates (Gehrels and Woodworth 2013; Kemp et al., 2014; Lambeck et al., 2014). A regional compilation of modern sea level records shows the transition from late Holocene slow background rates, to acceleration occurring around 1925 (Gehrels and Woodworth 2013). Recent SLR rates observed by satellite and corrected for vertical land movements, adjusted for bias drift in the sea level trend and tide gauge data, and GIA, show a further increase in the 19th Century rates. The precision of the altimetric data shows that sea level rose at a rate of 2.6 ± 0.4 mm / year from 1993 to 2014 (Watson et al., 2015). The records used to populate the modern sea level data base also include correction for other causes of relative sea level change including tectonics, sediment compaction and isostasy which cause regional and local variation in sea level. Isostasy is the major cause of the regional patterns of variation in relative sea level change (Lambeck et al., 2014).

2.2.4 Isostasy

At the end of an ice age, ice melting results in spatially variable relative sea level due to changes in the deformation of Earth's surface under time dependent varying ice and water loads. Time dependence refers to the evolution of surface loading of the Earth's crust over time, and the consequential isostatic response of Earth to this loading (Whitehouse, 2009; Lambeck, 2002; Lambeck et al., 2014). Concurrently the gravitational potential of the Earth – ocean – ice system changes, causing a combined contribution to relative sea level; termed the glacio – hydro isostatic effect (Lambeck and Chappell, 2001). Isostasy refers to the gravitational equilibrium between the Earth's lithospheric crust and the mantle it overlies, so loading of the crust over time causes deformation of the mantle (Whitehouse, 2009) resulting in a change in the relative position of the Earth's surface to the ocean.

The material properties of the Earth's crust and mantle, including their relative depth and density, are the governing parameters of the glacial isostatic adjustment system dynamics at any location (Whitehouse, 2009). Continental crust is deeper than oceanic crust and has higher concentrations of lighter elements such as aluminium and silicon. As a consequence, continental crust is less dense than ocean crust although is more laterally heterogeneous in depth and density. While oceanic crust is more consistent than continental crust, it has higher concentrations of magnesium and iron which results in a greater density. Oceanic crust behaves elastically under surface loading, meaning that it will deform and return to its original state after the force is relieved.

Beneath both types of crust, a transition to mantle occurs, marked by a reduction in silica content to denser, ultramafic materials. For any location or time, depending on the interaction between these material properties and the duration and pressure of surface loading, the lithosphere can behave elastically in that it deforms and rebounds viscously, or visco-elastically causing instantaneous, permanent deformation. The principle response of the lithosphere to glacio – isostatic adjustment (GIA) related surface loading is elastic deformation. The mantle response is viscoelastic, as it compensates for the lithospheric deformation and concomitantly continues to drive the isostatic response (Whitehouse, 2009).

The relative influence of the glacio isostatic effect is greatest in previously glaciated (near field) locations, whereby the rate of crustal uplift with unloading exceeds the rate of eustatic sea level rise and thus, relative sea level falls (Milne et al., 2006; Whitehouse, 2009). This is evidenced in figure 2.1 which shows the relative sea level curve for the Angerman River in the Gulf of Bothnia.

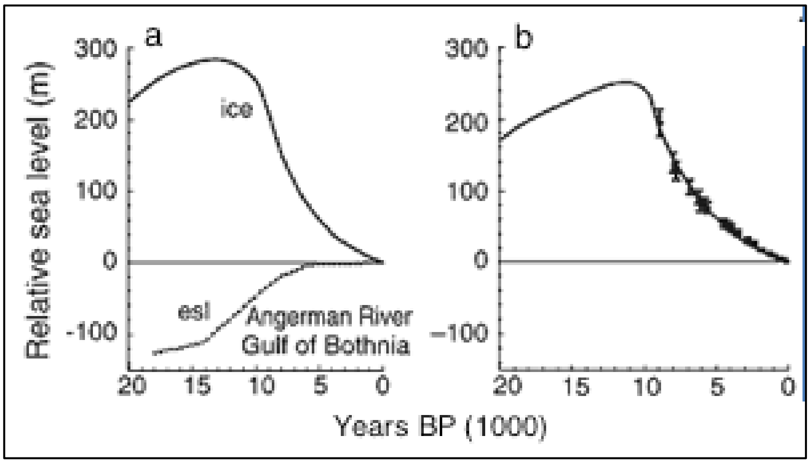


Figure 2-1 Relative sea level change for the Angerman River, Sweden a tectonically stable, near field location showing that while eustatic sea level rises with ice volume decline (a), relative sea level fall occurs (b), as the continental lithosphere rises at rates similar to glacial unloading. Source: Lambeck, 2002.

Variability occurs with distance from the ice sheet centre, because during glaciation great pressure is exerted on the crust by the glacier, causing the viscous mantle material to spread and bulge at the margins variably (Gehrels and Long, 2008). As the ice sheet melts the bulge subsides and continues to do so after complete ice melt. It is observed as slow relative sea level rise (Lambeck and Chappell, 2001), such as occurred at Upper Firth of Forth Scotland after 8,000 yr BP (Lambeck, 2002) (see figure 2-2). This pattern of sea level rise has been recorded across Great Britain where a vast number of sea level index points (SLIPs) have enabled continuous improvement in isostatic adjustment models and observed and predicted trends that are in increasingly good agreement (Shennan et al., 2006).

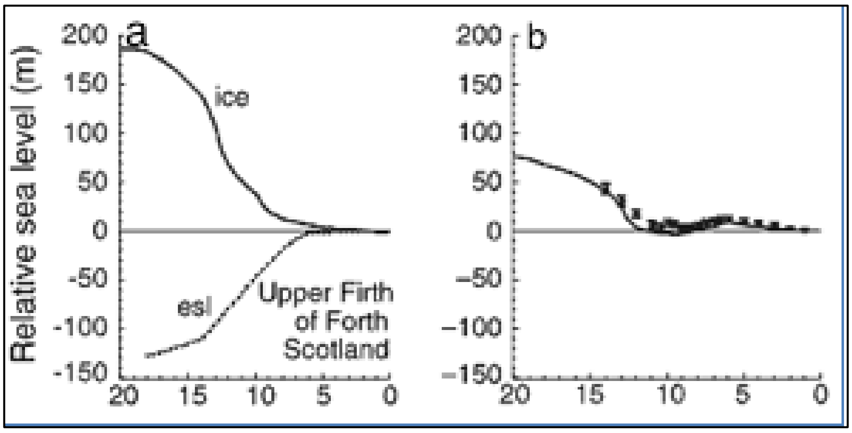


Figure 2-2 Relative sea level change for Upper Firth of Forth Scotland a tectonically stable quasi stable near field location showing eustatic sea level rise with ice melt (a), and b) slow relative sea level rise after 8000 years BP as the bulge at the ice sheets margin subsides after the ice melt. Source: Lambeck, 2002.

Remote from former ice sheets, particularly along continental margins, the dominant forcing factor is the hydro-isostatic effect. Water loading and consequent subsidence of the ocean floor cause adjustment to relative sea level (Lambeck and Chappell, 2001). These processes are first observed as relative sea level rise, as the oceans fill, followed by sea level fall as the sea floor and the coastal margin subsides. This far field pattern of relative sea level change has been observed in northern Australia (see figure 2-3).

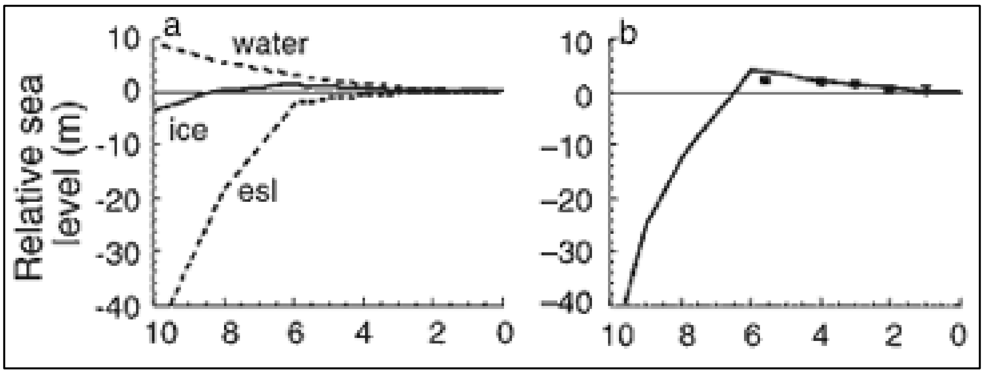


Figure 2-3 Relative sea level change for Karumba, Gulf of Carpentaria, Australia is consistent with modelled relative sea level change for a continental margin, far field site where the Earth adjusts to the water load added earlier with a small highstand followed by a sea level fall. Source: Lambeck, 2002.

The degree to which this pattern occurs is dependent on the distance from the melt water, and the dimensions of the ocean basin (Lambeck and Chappell, 2001). Holocene sea level history of Australia is consistent with this far field model.

Local sea level histories are critical inputs to regional GIA models (Gehrels and Long, 2008; Engelhart et al. 2011a; Kemp et al., 2014; Lambeck et al., 2014). In turn GIA models can predict shoreline position over time, to inform where gaps in sea level history occur (Whitehouse, 2009). For example, Shennan and Horton (2002) compiled 1097 SLIPs from 2212 records, using a validation process that 'filtered' potential errors using 70 fields of information, constraining them to an elevation at or above mean high spring tide level. The compilation of index points was compared with isostatic adjustment modelling to elucidate mid to late Holocene tidal change by accounting for the melting ice sheet that had covered most of Great Britain around 18 000 years BP. The study showed how differential crustal rebound around the isles resulted in very different relative sea level changes for different parts of its coastline (Shennan and Horton, 2002).

More recently, a sea level history for the United States of America's Atlantic coastline used the same approach of compiling a database of existing SLIPs as described above to develop a GIA model (Engelhart et al., 2009; Engelhart et al., 2011b). It filtered existing SLIPs from sea level specific studies and selected only those from the sedimentary record contained within saltmarshes as accurate (see section 2.5). All other dated samples were presented as "limiting dates" as the environment of deposition (elevation relevant to MSL), could not be assigned

Although the density of SLIPs established using accurate techniques of sea level reconstruction along the coastline are relatively sparse in Australia, compared to the areas such as the Atlantic Coast of America (Engelhart and Horton, 2012) and the British Isles (Shennan et al., 2012), there are indications that the far field model and the glacio hydro-isostatic adjustment model may apply to Australia (Nakada and Lambeck, 1989; Lewis et al., 2013). Figure 2-4 shows the compilation of SLIPs for far field locations around the globe that illustrate Earth's rheological response to glacial meltwater contribution from the Last Glacial Maximum to the Holocene Epoch (Lambeck et al., 2014).

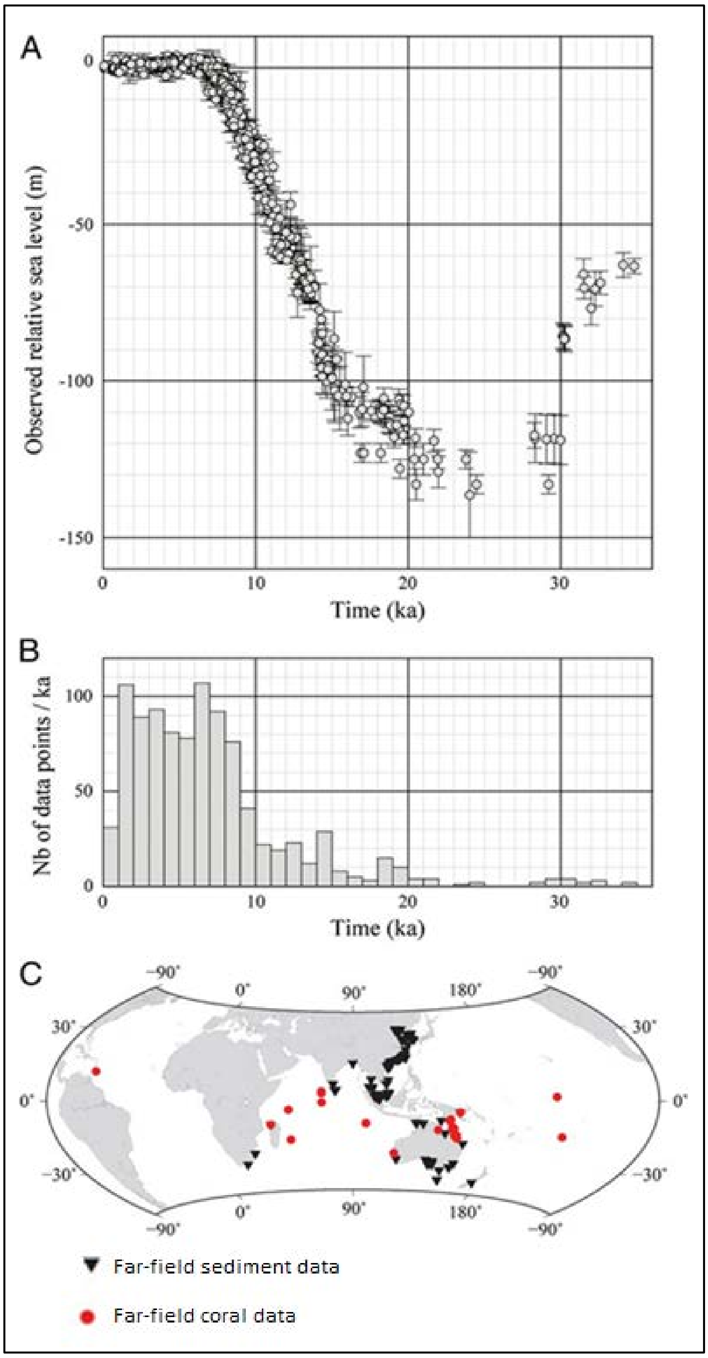


Figure 2-4 The glacial period sea level curve (A), compiled from far field locations shown in C. B, shows the number of data points used to construct the curve illustrating the increasing confidence in the data past after 10 000 yr BP. The curve shows the rapid fall in sea level at the last glacial maximum (from around 30 000 to 28 000 yr BP), followed by a period of stability to 20 000 yr BP, after which sea levels increased as the land based glaciers began to melt. The most rapid rate of sea level rise occurred between around 16 000 to 8000 yr BP during the main phase of deglaciation. (Source: Modified from Lambeck et al., 2014).

2.3 Australian sea level research

Lewis et al. (2013), in their review of sea level research in Australia illustrate the highly diverse nature of the Australian coast (Thom and Short, 2006), which extends from tropical to cool temperate environments from 9° to 42° latitude. This environmental diversity makes correlations between sites problematic (Belperio et al., 2002; Woodroffe and Horton, 2005, 2006; Lewis et al., 2013). No one indicator occurs over an ecological range at any one site spanning a sufficient time interval to provide evidence for a continuous sea level history (Lewis et al., 2013). Because Australian sea levels follow the far field glacio - hydro isostatic adjustment trajectory, much of the past record is below present mean sea level (Thom and Short, 2006), and much of the unconsolidated sedimentary coast lacks the sensitivity to record changes in geomorphic processes.

Thom and Short (2006) suggested that the extensive coarse sediment nature of the Australian coastline has encouraged a uniquely morpho-stratigraphic approach to research, which seeks to understand the morphodynamic relationship between coastline and sea level change. Estuaries have been key sites for examining landform evolution and sea level change in Australia. Sloss et al. (2007), compiled a sea level data base for south eastern Australia from work that commenced in the 1970s and 80s (Jones et al., 1979; Chapman et al., 1982; Thom and Roy 1983, 1985; Young et al., 1993) The compilation was drawn from sea level index points (SLIPs) obtained from drill cores of depositional sequences from estuaries, back barriers and the inner continental shelf. It identified that transgressive processes were still occurring between 9,000 – 7,000 yr BP and that sea level reached its present position around 6,500 yr BP; also that a stillstand was apparent at around 3,000 yr BP. However, while evidence of a > 1 m high stand was provided, considerable variability in the record prevented an understanding of height or duration. This period remained an “uncertainty zone” (Roy and Thom, 1984) until it was clarified by Sloss et al., (2007) who employed the same morphostratigraphic techniques with extensive sampling within estuaries in the same region, with the addition of improved dating methods for new and existing SLIPs

The additional work identified an estuarine transgressive sand sheet from which abundant intertidal mollusc macrofossils provided dating material (Sloss et al., 2006a, 2007). New SLIPs were limited to dates from fossils of known modern ecological range and taken from facies for which accurate descriptions existed and the stratigraphic relationship to present mean sea level understood. This data set significantly improved resolution of the timing and magnitude of mid to late Holocene sea level change for the New South Wales coastline of south eastern Australia. Modern sea level was attained between 7,900 and 7,700 cal. yr BP. A highstand of +1 to 1.5 m occurred between 7,700 and 7,400 cal. yr BP to 2,000 cal. yr BP, after which the sea fell to its present level (Sloss et al., 2007). Later research comparative sites in the region continued to support the geochronological framework (Sloss et al., 2010).

Other Australian studies of coastal evolution in estuaries from which sea level history can be inferred have been undertaken in South Australia, where good regional resolution of the mid Holocene highstand was detected (Belperio, 1993; Belperio et al., 2002). Northern Australia’s South Alligator River, Adelaide and Mary Rivers provided a record of marine influence in presently brackish environments (Woodroffe et al., 1993). In Queensland, sea level change is deduced from coral micro atolls and fixed marine indicators. Sea level indications in Tasmania are few, with no precise indicators of past sea level earlier than 650 yr BP generated by Gehrels et al., (2012). Table 2.1 summarises the review by Lewis et al. (2013), of research on Holocene sea level change for Australia, with reference to the primary sources.

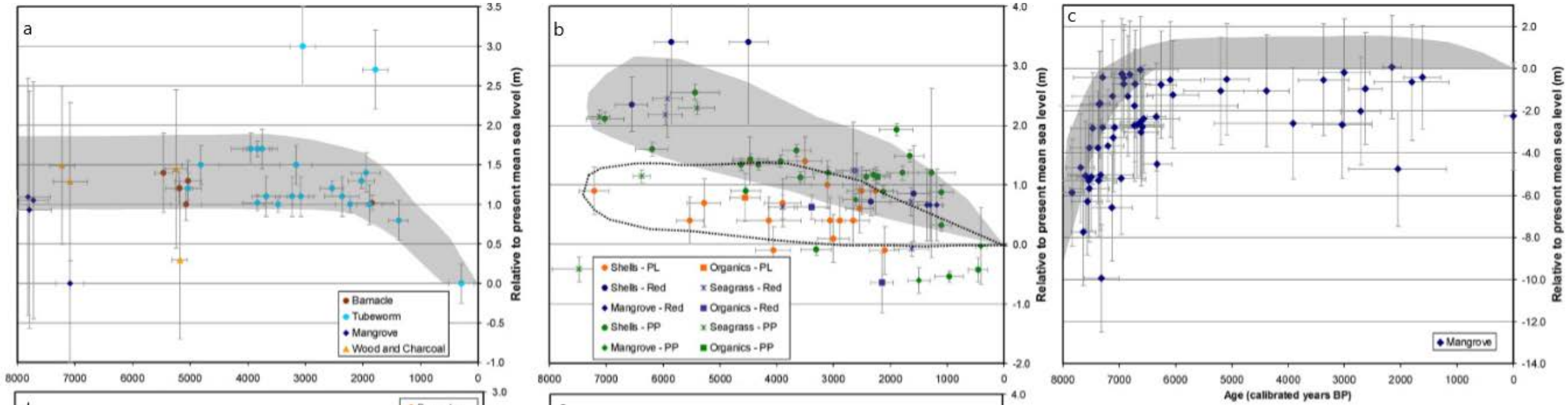
Table 2-1 A Summary of sea level history prepared from the most accurate sea level indicators for eastern Australia, as per Lewis et al. (2013). Collectively, evaluation of the precision and accuracy of these studies provide a data set to inform where future research efforts should improve resolution of the timing and magnitude of sea level change, and to fill gaps in the spatial record. It concludes that while accounting for the likely range of uncertainty inherent with different indicators and broad geographic extent, the current data set illustrates variation in the timing and magnitude of sea level change. This is most obvious for the mid Holocene highstand between regions, as a consequence of the differential glacio eustatic and-hydro isostatic adjustment processes, antecedent geomorphology, shelf characteristics and oceanic and climate controls. Note column entries are only filled where the information is available, showing how regional sea level history is largely pieced together from different locations and proxy indicators

Location & authors	**Culmination cal. yr BP**	**Highstand**		**Regression**	**Interpretation** (by Lewis et al., 2013)
		Magnitude(m)	**Timing (cal. yr BP)**		
Northern Australia					
South Alligator River (Woodroffe et al., 1985, 1987, 1989, 1993)	7400 ±200	Not detected.			Some pollen evidence of transition from mangrove to grasses transgression and fall (Woodroffe et al., 1987)
Karumba, Gulf of Carpentaria (Rhodes 1980; Chappell et al., 1982; Rhodes et al., 1982)		+2.5*	6400 ± 90	After 1000 yr BP, a smooth fall to PMSL.	Not precise indicators with vertical errors of ca ± 2.5 m.
McArthur River (Nott, 1996)		+1-2	ca1500 - 3300		Evidence of highstand on beachrock, bearing some consistency with Karumba as above.
Queensland					
Central Great Barrier Reef, (Grindrod and Rhodes 1984)	8170±240 within 3m of present*				Consistent with Woodroffe's conclusion of culmination between 8000 – 6200 cal. yr BP. (2009).
Magnetic Island, (Higley, 2000; Lewis et al. (2008)		+1*	7380±240	Smooth pronounced fall after 2000 yr BP.	Highstand timing and magnitude inferred from synthesis by Lewis et al., 2008 of SLIPs from Yu and Zhao, 2010; Zwartz, 1995.

Central north Great Barrier Reef, Chappell et al., 1983; Yu and Zhao, 2010		~ + 1* +1.3 – 1.5	~ 6000 7012±22	Smooth fall to present.	Coral micro atoll and encrusting fixed indicators data considered most reliable.
Beaman et al. 1994; Higley, 2000; Lewis et al., 2008		+1.6 - +1.7	6280-5720		Correlates well with those above for Queensland.
South eastern Australia					
Jones et al., 1979	7900 - 7700				Based on one ^{14}C date
Sloss et al., 2004, 2007.		+1.0 – 1.5	7700 - 7400	Gradual fall to present from 2000 cal. yr BP.	Time averaged compilation of ^{14}C and AAR ages of transgressive facies, plus revised published data.
South Australia					
Redcliff , Spencer Gulf (Belperio et al., 1984; Belperio et al., 2002)	8000 – 7500	+3 ± < 1	6400	Smooth fall to present through mid to late Holocene	Large variability across sites due to hydro-isostatic adjustment consistent with GHIA model of Nakada and Lambeck (1989). Tidal frame transitions provided most precise indicators in the region.
Port Lincoln (Belperio et al., 2002)		+1 ± 0.4			

- *** Vertical errors not available**

There is agreement between the observed relative sea level change (Table 2.1) and the glacio - hydro isostatic adjustment model predictions of Lambeck and Nakada (1990). Consistent with far field locations, the sea level signal is dominated by the glacio-eustatic sea level term in which its magnitude is equivalent to the volume of melt water contributed from the diminishing global ice sources (Lambeck and Nekada, 1990). The geographic variation from northern Australia around the east coast to South Australia is consistent with the variability likely with the water load on variable continental shelf width (Lambeck and Nakada, 1990; Lewis et al., 2013), including the predicted progression in highstand magnitude from Tasmania to Northern Australia (Lambeck and Nekada, 1990). Figure 2-5 compiles the sea level envelopes for each region, that Lewis et al. (2013) generated from records that had horizontal error values, or that had enough information to evaluate horizontal error since their first publication. However, no such envelope was generated for Tasmania, due to the lack of data points to indicate change over time, and for the data that was available, there was a lack of information to evaluate and assign horizontal error margins. Chapter Four provides a review of studies that infer past sea levels for Tasmania.



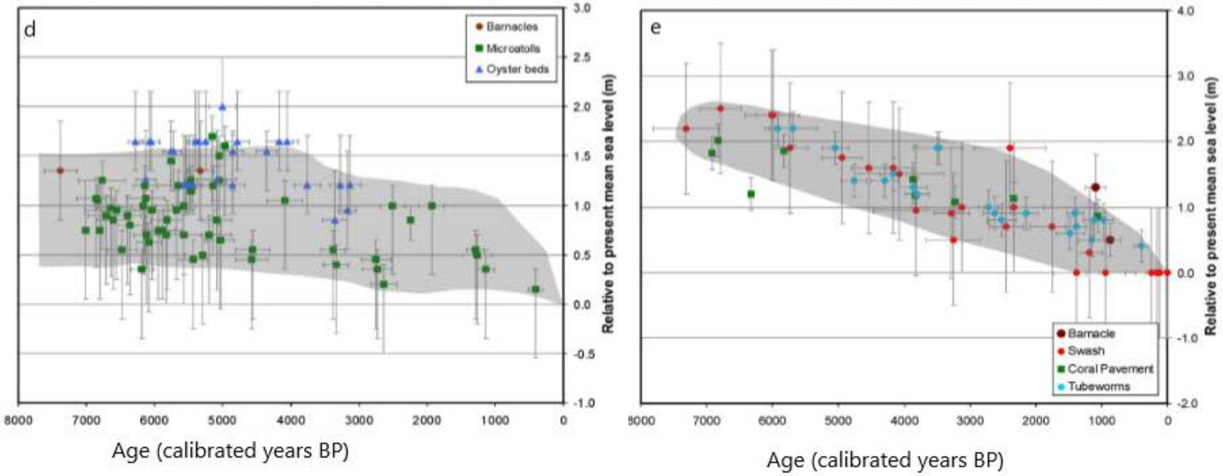


Figure 2-5 Sea level envelopes for a) New South Wales, b) South Australia - Spencer Gulf c) Nothern Territory, d) Queensland and e) Western Australia. The sea level trends since 8000 yr BP show considerable differences in the timing of present sea level attainment, and the magnitude and duration of the mid Holocene sea level highstand. The compilation also shows the variety of species used to indicate past sea levels across regions.

Source: Modified from Lewis et al, (2013)

2.4 Palaeo-reconstruction of past sea levels in estuaries

2.4.1 Sea level index points from the stratigraphic record

The results of sea level change studies are described graphically as a series of SLIPs, plotted as time against elevation relative to mean sea level (msl). Horton and Edwards (2006) refer to such charts as age-altitude plots. Due to the curvilinear trend of most results from studies that reconstruct sea levels for the Holocene epoch, they are termed sea level curves. They can be developed using simple techniques for any location for which there is datable material that can be matched to a quantifiable elevation relative to mean sea level (van der Plassche 1986; Gehrels, 2002; Horton and Edwards, 2006).

In practical terms, judicious sampling of the stratigraphic record for dating is required to produce a meaningful curve or fill gaps in chronology, but this requires substantial understanding of the likely locations that will provide such evidence. Estuarine structures, and variation in boundary conditions and forcing factors, need to be understood to be able to interpret the sedimentary record (Sloss et al., 2007).

The sedimentary record is retrievable using coring techniques that preserve facies sequences as they were deposited. Most authors, (e.g. Boggs, 1987; Doyle et al., 1995) point to a unifying definition of a facies after Gressly, (1838, in Cross and Homewood, 1997), in that a facies is the sum total of the characteristics that enable its interpretation. The broadest interpretation is that a facies is a stratigraphic unit that is distinguished from other units by lithological characteristics (lithofacies), or by fossil content (biofacies) (Boggs, 1987). Facies can be mapped vertically and laterally, thus enabling the development of a spatio-temporal model that represents the process and landforms associated with sea level history.

Facies sequence models are simplifications of the real world, created to assist our understanding of complex situations (Reading, 1986; Boggs, 1987). One of the first estuarine evolution models that described the process of estuarine infill during the Holocene sea level rise and subsequent still stand, was developed by Dalrymple, Zaitlin and Boyd (1992). Their model describes lateral and vertical facies sequences for regions of the estuary, divided on the basis of hydrological energy levels, thus defining both spatial structure and through the facies sequences, its temporal span. It has been widely used as an inductive model of the evolution of micro-tidal, wave dominated, drowned river valley estuaries built on sedimentary coasts, particularly for the south eastern Australian coastline. For the wave dominated energy regime, it used the facies architecture developed for microtidal, barrier estuary evolution in New South Wales estuaries (Roy et al., 1980; Roy and Thom, 1984). It remains widely used (Heap and Nichol 1997, Dabrio et al., 2000; Wilson et al., 2007; Dawson et al., 2012) and is the basis of the most recent and comprehensive classification of coastal systems around Australia (Ryan et al., 2003). Since its publication, further detail has been added for microtidal estuaries in tectonically stable regions (Sloss et al., 2004, 2006a, b, 2010) and a model accounting for neotectonism (Wilson et al., 2007), both described in Section 2.5.1 and 2.5.3 respectively

2.4.2 Using inductive models to build an actualistic model

Inductive models are used to aid understanding of complex processes and phenomena by retaining the fundamental features of the real world and their interactions (Reading, 1986). They are generalised because the site specific features of variability are filtered out. Actualistic models retain site specific features, so that the landform evolution reconstruction is as realistic as possible, and the processes and phenomena specific to a site that created the landform can be understood.

The primary aim of an inductive model is to aid interpretation and development of actual facies models by enabling the formulation of an initial working hypothesis from which predictions can be made as to the most appropriate location to find facies sequences (Reading, 1986). Comparison of the structure of a real world estuary in plan form, with a model, should enable palaeogeographical interpretation of the site, so that the past environments can be deduced from modern ones.

The process requires understanding of the basic principles of sediment transport and deposition. Variations to inductive models can occur as a consequence of the amount, type and source of sediment available for infill and the antecedent topography that affects the energy gradient that variably erodes, transports and deposits the sediment (Dalrymple et al., 1992). All of these factors interact with tidal range, climate, catchment size and geology, and the rate of relative sea level change. When it is time to examine and interpret facies characteristics made evident in sediment cores, the variations these create become particularly relevant. Figure 2-6 shows the process of enquiry for sea level studies in estuaries, using the available inductive models and knowledge of processes in the region as a framework for producing an actualistic model of sea level history.

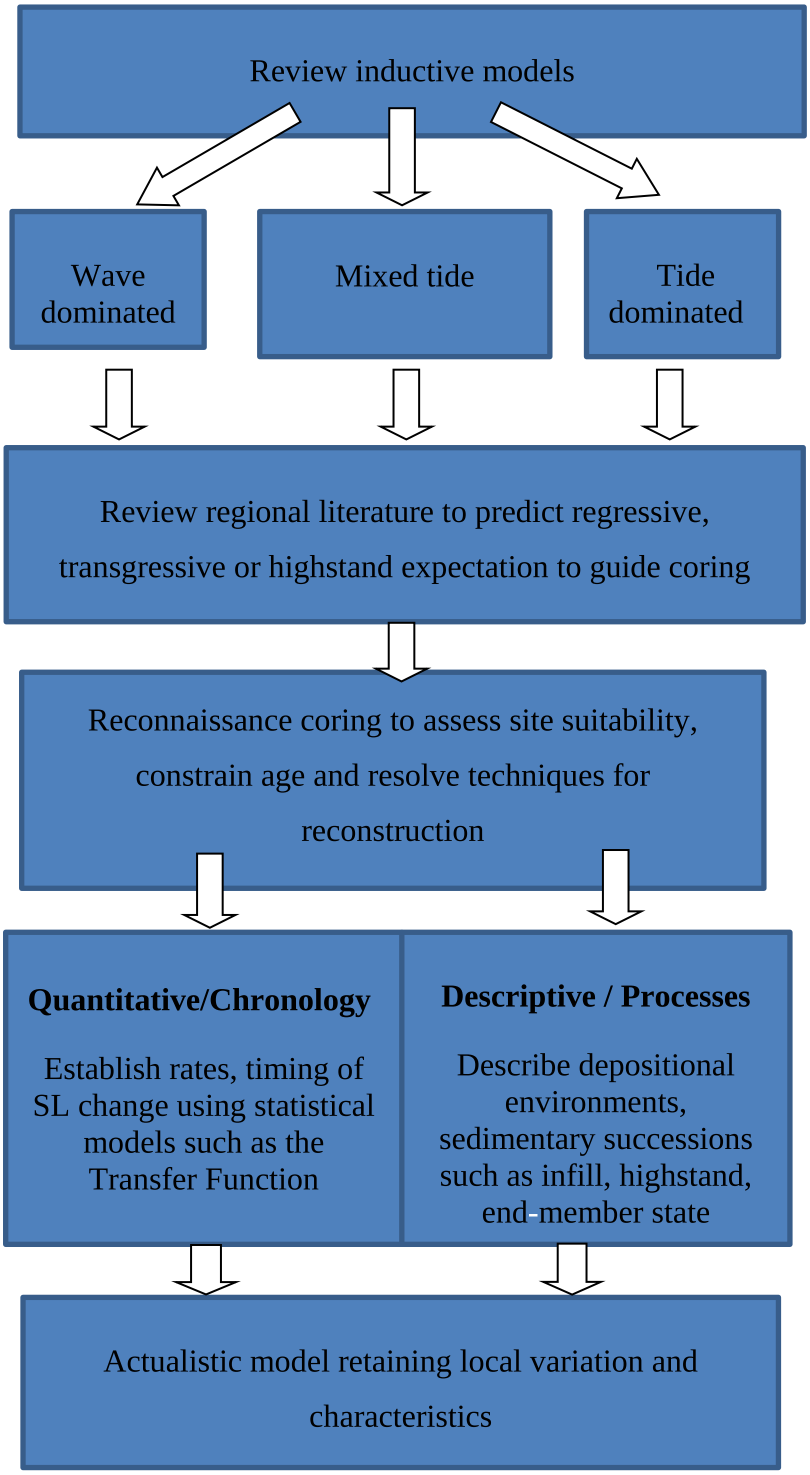


Figure 2-6 A procedure for sea level studies in estuaries.

2.4.3 Desktop review of estuarine structure as per wave, tide and mixed wave / tide dominance

Basic estuarine classification can be done using aerial photographic analysis. There are fundamental differences between the wave and tide dominated estuaries but many are inter-gradational in some way and are termed mixed energy estuaries (Perillo, 1995). However, in all but a few cases (Heap and Nichol, 1997), the tripartite zonation described by Dalrymple et al. (1992) is relevant to all estuaries. The interactions between fluvial and marine processes are predictable in their longitudinal variation in intensity. Dalrymple et al. (1992) emphasise that variations are possible due to the local or regional differences in forcing factors and boundary conditions. Unless otherwise stated, the following summary of the basic models is summarised from Dalrymple et al. (1992).

2.4.3.1 Tide dominated estuaries

Meso to macro tidal ranges encourage tide dominated estuaries as a consequence of the great energy large tides maintain. However, tide dominance is known to occur where tidal ranges are smaller if the wave action is limited and or the tidal prism is large (Perillo, 1995). When the tidal energy is greater than the wave energy at the mouth, elongated sand bars are created. These dissipate wave energy that further decreases up the estuary. High tidal energy creates a funnel shaped geometry which causes a constriction of the flood tidal prism, causing it to increase its speed into the estuary until frictional dissipation exceeds the effect of the convergence and the tidal energy decreases, reaching zero at the tidal limit.

The point at which fluvial and tidal energies are equal is the location of the minimum in the total energy curve. This creates a muddy depocentre in the middle reaches that is more pronounced than in wave dominated estuaries because the tides exert energy further up the estuary than waves. Hence the tripartite zonation in tide-dominated estuaries is not as pronounced and sands can occur in the tidal channels throughout their length. As a result, muddy sediments can be confined to tidal flats and marshes along the sides of the estuary. The geomorphology of each zone is thus:

a) a marine sand body consisting of elongate bars and broad sand flats that pass headward into a low sinuosity (straight), single channel. Net transport is headward,

b) a central zone consisting of tight meanders, bedload transport by flood tide and river currents is roughly equal in the long term, and tidal channels may be of fine sand, and

c) an inner river dominated zone that has a single, low sinuosity (straight channel).

In Australia, the best studied tide dominated estuaries have been in the macrotidal settings of tropical northern Queensland and the Northern Territory. Woodroffe et al. (1993) described estuarine plains that have developed since sea level stabilised and fresh water clays are now being deposited over the estuarine sediment. However, variation in the rate of deposition was observed. This variation is a product of the east – west gradient in coastal sedimentation that correlates with sediment availability from longshore drift.

2.4.3.2 Wave dominated estuaries

Wave dominated estuaries are most likely to occur in micro to mesotidal ranges. Because wave energy does not penetrate as far up the estuary as tidal energy, a low energy central deposition zone is present. Longitudinal energy variation creates a coarse – fine – coarse lateral lithofacies distribution. At the mouth is a marine sand body that accumulates from longshore drift, and wave energy deposition. Eventually a barrier forms, and then acts to reduce wave energy entering the estuary. Tidal influence is still present inside the mouth and creates a flood tide delta, as marine sediment is transported landward.

The major difference between wave- and tide-dominated estuaries in the central basin is that wave dominated estuaries display a straight, to meandering, to straight channel morphology. The outer straight reach is tidally dominated and the net sediment transport is headward due to strong tidal currents. The channel contains alternate bank-attached bars and some mid channel bars. The inner straight reach also contains similar bar types, but here net sediment transport is downstream due to the long term dominance of river flow over tidal currents. The region between the two straight stretches contains tight meanders with symmetrical point bars.

The key features of the tripartite zonation of a wave dominated estuary include:

a) a marine sand body such as a barrier, washover, tidal inlet and tidal delta deposits,

b) a central basin consisting of fine grain muddy sediment. It acts as a prodelta region if there is an open lagoon, and fine grained muds accumulate there. In nearly filled lagoons extensive salt marshes exist because of reduced energy conditions, and

c) a bay head delta influenced by salt and freshwater, and fluvial sediment delivery.

In New South Wales (NSW) southeastern Australia, micro tidal wave dominated barrier estuaries are common because of sufficient sediment from offshore, and because rivers rarely exert much influence due to small catchment sizes and, or, low rainfall (Sloss et al., 2006a). Figure 2.7 shows the evolutionary stages of barrier estuaries with deposition of sediment brought offshore during the marine transgression and subsequent Holocene sea level highstand.

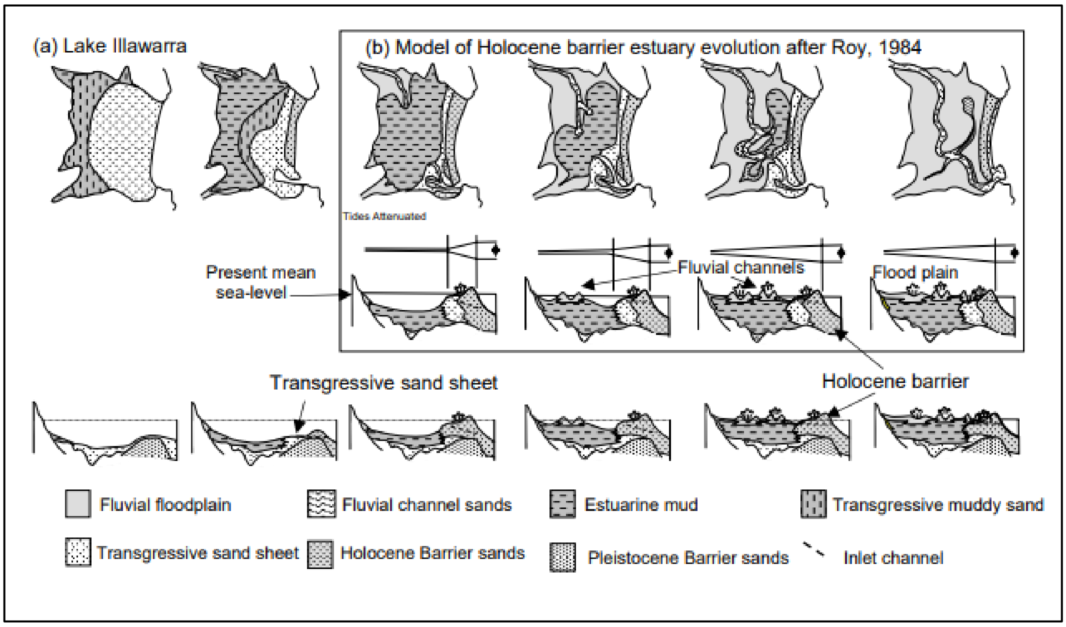


Figure 2.7 Barrier estuary evolution on a tectonically stable coastline showing the stages of barrier evolution where marine sediment supply is sufficient and river influence low.

Source: Sloss et al, 2004.

2.4.3.3 Mixed energy estuaries

An example of a mixed energy estuary that utilises the wave and tide dominated inductive models to aid interpretation and development of an actualistic model, is the Gironde Estuary off the Atlantic Coast on the Bay of Biscay, in south western France (Allen and Posmentier, 1993). Here the Atlantic coastline experiences high wave energy that enters an unrestricted estuary mouth that is ca. 18 km wide. The wave action produces strong littoral drift that delivers sediment to the estuary mouth and forms a barrier consistent with the wave-dominated model. Also consistent is its tripartite zonation; these estuaries having a deep central depocentre and a clear fluvial dominated zone toward the head of the estuary. However, rather than taking the typical form of a bay head delta as per the wave-dominated estuary, there are tidal sand bars that transition into tidal estuarine point bars as the tidal channel meanders through the incised valley to the river-dominated zone. The macro - meso tidal regime of 5 m spring tides enables the tidal current to extend 130 km inland during low river discharge and 100 - 115 km during river floods (Allen and Posmentier, 1993), thus tidal energy is a dominant forcing factor.

Another variation was described by Heap and Nichol (1997), who also used the inductive models to interpret the evolution of a landward depocentre in the Weiti River, New Zealand. Although their model was broadly consistent with the wave-dominated inductive model, the narrow and steep sided valley topography excluded a broad central basin to serve as accommodation space for the transgressing sea. Thus, the typical low energy depositional zone for infill is absent. Other proposed causes include rapid fluvial transport into the landward depocentre zone during transgression and still stand, with biostratigraphic evidence indicating supra-tidal to intertidal environments. The authors also suggest that a lack of a barrier in the late Holocene prevented establishment of a central basin in the landward depocentre (Heap and Nichol, 1997).

2.5 Sea level sequences and their facies

All of the structural models described in the previous section were the result of sea level transgression. They are the result of the evolution of modern estuaries that have occurred due to the processes of sediment infill during the Last Postglacial Marine Transgression. The models rely on simple representations of the vertical stratigraphy of estuarine infill. However, it is important to recognise that there are different sequences of sea level that will be identifiable throughout the stratigraphic record in estuaries, including combinations of transgression, regression and sea level still stand. These processes should be evident, but facies architecture becomes more complicated vertically. The following section describes the different types of facies for each sea level trajectory, including the transgression, highstand, regression and stillstand.

2.5.1 Transgressive and highstand facies

Sea level transgression refers to the progressive inundation of the land surface by the tides. The transgressive facies sequences exemplified for the wave-dominated micro-tidal drowned river estuary model developed by Roy et al. (1984) and described by Dalrymple et al. (1992) has been further developed by Sloss et al. (2004; 2006a, b; 2010) whose actualistic models illustrate the influence of sediment availability and antecedent topography on landform construction.

At Lake Illawarra in NSW, a barrier estuary, five stages of estuarine evolution linked to sea level change are evident from the spatial and temporal organization of facies (Sloss et al., 2004): Commencing with Stage 1) sea level lowstand fluvial valley incision during the last glacial phase of the late Pleistocene, 2) Holocene sea level transgression indicated by a marine mollusc rich transgressive sand sheet unconformably overlying the late Pleistocene surface, 3) sand barrier formation during the + 1 to 2 m mid Holocene highstand that restricted marine influence to create a back barrier lagoon environment, 4) late Holocene sea level regression, causing shifting tidal inlets and fluvial delta progradation over the older estuarine sequence , and 5) estuarine infilling with increased fluvial delta progradation and declining marine sediment deposition with sea level stabilization.

A similar sequence at Lake Burrill NSW supported the evolutionary pathway associated with sea level transgression which culminated in a sea level highstand of around + 1 – 2 m, ca. 7800 years ago, lasting until ca 3000 years ago before it fell to present levels, stabilizing around 2000 years ago (Sloss et al., 2006b). The same approach was undertaken at Lake Conjola, which also formed a mid Holocene back barrier lagoon environment, overlying a near basin transgressive sand sheet, confirming the evolutionary pathway of wave dominated barrier estuaries in the region (Sloss et al., 2010).

The compilation of models from NSW barrier estuaries provides an inductive, chronostratigraphic framework for descriptions of Holocene sedimentary sequences in tectonically stable drowned river valley estuaries. Holocene barriers formed in regions that experienced sea level transgressive to highstand followed by regression. The chronostratigraphic framework allows sea level history to be deciphered from geomorphological changes.

2.5.2 Regressive models

Regression only processes are not as easily identified through facies architecture in estuaries as transgressive sequences. Sea level regressions during the Holocene have been deduced from higher than present sea level landward deposits (Doyle et al., 1995), and from the palaeo-geomorphic features described above in 2.5.1.

Heap and Nichol (1997) identified the full history for the Weiti River Estuary in New Zealand and were able to draw broad parallels with the inductive model of Dalrymple et al. (1992). Although the focus of their study was the influence that limited accommodation space and valley shape have on facies organisation, they identified eight lithofacies, arranged in two distinct depocentres. The first lithofacies is the marine influenced zone of sand accumulation, representing the transgressive to highstand open bay deposits. These are overlain by regressive tidal flat and shelly beach ridge deposits which began to develop around 3,000 years BP. The landward depocentre is dominated by tidal / fluvial transgressive, to a highstand bay head delta complex that includes point bar, channel lag and salt marsh deposits. The diagnostic facies for marine transgression that is comparable with Dalrymple’s model is a transgressive bay head delta.

The stratigraphic arrangement of facies varies from Dalrymple et al. (1992) model by the absence of a transgressive central basin deposit overlain on a prograding bay head delta. The likely reason for the significant departure from Dalrymple’s and subsequent models is interpreted to have occurred as a consequence of the delay in the formation of the seaward barrier. This would otherwise have attenuated energy and subsequently reduced sediment size typical of the central basin being transported and deposited.

Other useful models of sea level regressive architecture describe abandoned successive beach ridges (Harvey, 2006; Ribolini et al., 2011), although most regressive models involve bay beaches rather than estuaries. Engels and Roberts (2005) present an actualistic model of beach ridges and their relationship to adjoining environments in a coarse sediment setting within the Fraser River Delta in Australia. The beach ridges were shown to have prograded over the previous tidal flat when sea levels stabilized at their maximum height of the highstand. Preservation of the ridges was interpreted as a feature consistent with a shift from transgressive processes, to stable conditions during the highstand. Ridges are now located well above present mean sea level.

It is important to recognise the difference between a regressive sequence developed as a consequence of estuarine evolution through the process of progradation and a regressive shoreline occurring as consequence of relative sea level fall because the diagnostic facies features will look similar for both. Posmentier and Allen (1999) differentiate normal and forced regression; the former being shoreline regression due to sediment supply outpacing sea level rise. In this instance, the interaction between processes causes a seaward migration of the littoral facies with rising relative sea level. Facies are characteristically progradational and aggradational in that they build seaward and vertically.

Forced regression occurs with sea level fall and the progradational facies are still present, but aggradation is not possible and is replaced with down-stepping stacking patterns as described by Catuneanu (2011). These models are useful and illustrate the general concept of regression sequences very well, although they have been developed for alluvial coastal plains and are thus less complex than regressive sequences in estuaries.

2.5.3 Tectonic models

Studies conducted in regions whose tectonic history is not well understood, should consider the possibility of encountering sequences that have been influenced by past tectonic movements. Wilson et al. (2007) reconstruct an incised valley infill sequence on the active, co-seismically uplifting, eastern coastline of New Zealand's North Island, (3.2 ± 0.8 mm / yr), and compare it to the original model of Dalrymple et al. (1992). Significantly the later model contains both the basal floodplain and the barrier / tidal inlet of Dalrymple, and thus captures the whole sequence of the Holocene valley infill in an outcrop exposure of raised Holocene terraces.

The Holocene marine terraces are offset by the Pakarae Fault (a normal fault). The study was made possible by the known rate of continuous tectonic uplift for the region. Palaeogeomorphological maps of river valley development were produced showing the evolution of the valley, and a sea level curve for the period 11,000 – 7,000 years BP was constructed. Sea level change during that time appear in the stratigraphic record as sea level falls, and observable horizontal bedding in the field allowed for the sampling, analysis and definition of biolithofacies and characteristic lithology, linking processes and past environments of deposition. Chronostratigraphic analysis enabled the timing and magnitude of tectonic events to be quantified. The resulting model is extremely useful and should be added to the list of globally applicable inductive models for actively co-seismic drowned river valley estuarine settings.

Another useful contribution to the field of palaeo sea level research in estuaries is the comparison provided by Wilson et al. (2007) of characteristic lithofacies between estuaries on the east coast of Australia (Sloss et al., 2004; 2006 a,b), with the Weiti River Estuary New Zealand (Wilson et al., 2007), both having very different geologies. The bay head delta in Australia is a transgressive, carbonate rich sand sheet, and because it lacked lithic components was most likely sourced from the continental shelf. However, New Zealand's younger geology provides ample terrestrial sediment, and this was suggested by Wilson et al., (2007) to ‘mask’ marine sediment inputs. Despite their relative geographic proximity, the two geologies cause very different diagnostic facies features, highlighting the importance of interpreting facies within the context of the site specific geologic and climatic settings. More broadly, the methods of Holocene sea level reconstruction used in Australia and New Zealand contrast with northern hemisphere reconstructions which focus on salt marsh environments. Understanding the biogeomorphic features of salt marshes peculiar to their climatic and geomorphic setting should similarly support the interpretation of sea level history contained within them.

2.6 Sea level research in salt marshes

Salt marshes have been studied extensively for the sedimentary record contained within them (van de Plassche, 1986; Kastler and Wiberg, 1996; Donnelly and Bertness, 2001; Haslett et al., 2001, 2003; Szornick, 2006, Gehrels et al., 2005, 2008, 2012; Engelhart et al., 2011a: Kemp et al., 2012, 2014; Barnett et al., 2016), because they only develop in quiescent areas, thus the record is likely to be continuous. Although they are locations of net deposition, high energy erosion events can occur, but are detectable in the form of unconformable contacts between facies. In this way, they provide a means of determining changes in sea level, according to lithological changes (Gehrels, 2002; Church et al., 2010). The techniques for employing ecological indicators such as foraminifera, diatoms and ostracods as finite indicators of past sea levels have undergone refinements (Wright et al., 2011; Kemp et al., 2011, 2015) and the taxonomic descriptions and biogeography of many species are available (van de Plassche, 1986; Gehrels, 2002, Hayward et al., 1999, 2011a). Researchers exploit knowledge of the highly interactive nature of the boundary conditions and forcing factors that create saltmarshes and their ecology, to determine the best indicators to suit the environment and time frame to be reconstructed and the required level of precision.

The nature of sea level reconstructions in estuaries, including salt marshes, range from descriptions of changes in species composition that reflect broad ecological change linked to sea level change (Cann and Bourman, 2000; Hayward et al., 2004), to highly quantitative studies such as transfer functions (Zong and Horton, 1999; Gehrels, 2000, 2002; Gehrels et al, , 2001,; 2008; 2012; Massey et al., 2006; Callard et al., 2011; Kemp et al., 2013). The latter are a mathematical expression of the relationship between the environment of deposition as indicated by microfossil species assemblages and mean sea level.

The resolution that transfer functions provide have been vital in understanding the nature and magnitude of sea level change over the late Holocene (Gehrels, 2002; Lambeck and Chappell, 2001; Lambeck et al., 2014), as well also detect the influence of the anthropogenically induced global mean sea level rise (Gehrels, 2012). A typical statistical procedure can be seen in figure 2-8, and is described in more detail in Chapter 5.

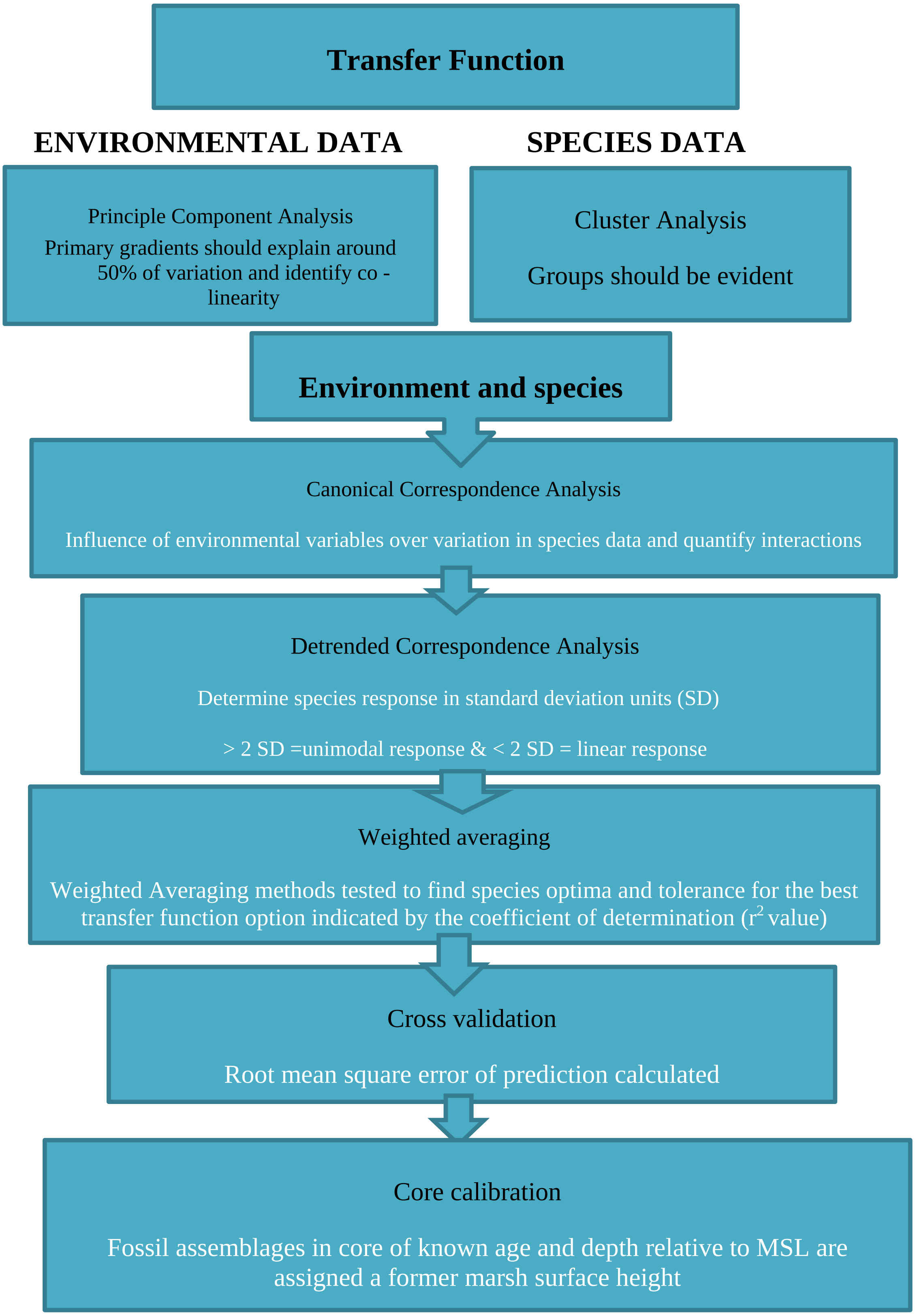


Figure 2-8 Typical procedure towards development of a transfer function.

Transfer functions can be built on a training set of elevation and species assemblages alone (Callard et al., 2011), although more reliable reconstruction is likely when as many environmental variables as practicable are included (Smol, 2002). This enables the influence of the variable of interest to be accurately assessed.

The majority of sea level reconstructions using salt marsh indicators have used the relationship between foraminiferal zones and their elevation relative to mean sea level as a proxy for tidal inundation period. Tidal inundation is the actual environmental variable of interest, as elevation alone exerts no direct influence over species (Haslett, 2002). However, tidal inundation period also influences other growth limiting factors such as soil salinity, pH, redox potential and moisture content, and these interact with climatic, topographic and geomorphological factors (Horton and Edwards, 2006)

The temporal span that a transfer function can reconstruct from salt marsh stratigraphy will depend on its age. The age and depth of salt marshes is variable according to the evolutionary stage of the estuary in which it exists. This is dependent on geomorphic setting, including distance from glacial outwash after deglaciation, and the ensuing climate (Pratolongo et al., 2009). These factors influence rates of sediment delivery to the estuary, sediment supply being critical to build and maintain saltmarshes during sea level rise. These factors, along with antecedent topography, collectively influence the evolutionary pathway of the salt marsh. It is important to understand these factors and their effect on salt marsh features pertinent to sea level change studies. For the purpose of comparison with what is known about the age and evolutionary pathway of Tasmanian saltmarshes, the following section details global scale variation, and describes the features of saltmarsh that make such environments conducive to sea level reconstruction research.

2.6.1 Global saltmarsh biogeomorphological groupings

Global saltmarsh groupings were first developed by Chapman (1960 & 1974) based on 9 biogeographic regions of community and species distributions across continents. Later, Adam (1990) grouped saltmarsh types according similarities at the genus level across regions, noting vicariant species across continents in temperate climates, such as from the genus *Juncus*, that occur commonly in the upper marsh, and *Sarcocornia* species that occur commonly in the pioneer zone and low marsh in the absence of *Spartina* spp. (Adam, 1990). Nine major groups reflect geographic, genus level and climate differences. European marshes were grouped by Dijkma (1984) and later by Pye and French (1993), who described seven marsh types based on geomorphic setting. The more recent grouping by Pratolongo et al., (2009), reverts to the continental scale ecological groupings but adds their geomorphic setting, including sea level history in recognition its influence over marsh development and ultimately biological zonation. For the purposes of recognising marsh features pertinent to palaeo-reconstruction of past sea levels and coastal evolution, this grouping is most useful.

The saltmarshes of the eastern Atlantic coastline of the U.S.A. and Canada are well studied (Pennings and Bertness, 2001) relative to southern hemisphere marshes, and understanding their biogeomorphology has contributed significantly to the development of palaeo reconstruction methods, following early work by Redfield and Rubin (1962) and Redfield (1972). The northern Atlantic group include the semi diurnal, meso to macro tidal marshes in the Bay of Fundy, a region dominated by barrier island geomorphology, built during the last 5,000 years of relatively stable sea level (Chapman 1974; Pratolongo et al., 2009). At Machiasport, Maine, marshes are dominated by *S. alterniflora* at the seaward edge at around -1.5 m relative to MSL (just below mean low tide level), and are inundated by up to 4 meters of tidal water daily. Behind this at higher elevations just above mean high water, *Spartina patens* dominates, above which diversity increases into the high marsh (Gehrels, 2002) Saltmarsh cores 1.2 m deep provided a 1000 year old record of sea level change. Further south, *S. alterniflora* dominates the low marsh of the New England group and is inundated daily (Pratolongo et al., 2009). In Connecticut, 2-3 meter *Spartina* spp. dominated saltmarsh deposits provided a 2200 year sea level history (Kemp et al., 2015). In this region, saltmarsh deposits range from shallow 60 cm thick marshes in recently filled sandy shallow lagoons, to around 10 meters in deep bays and estuaries, underlain by glacial outwash sediment (Bloom, 1964; Hill and Shearin, 1970; Stumpf, 1983). The latter record of the last 3,000 years showed continuous sea level rise of around 2.75 meters (Hill and Shearin, 1970). The most southerly, coastal plain group generally occupy shallow areas behind Pleistocene barrier islands or deltas and are similarly dominated by *S. alterniflora* at the seaward edge.

The marshes of Great Britain are also dominated by *Spartina* spp. at the seaward edge (Adnitt et al., 2007), but are considered “profoundly different” to US marshes (Horton and Edwards, 2006). As part of the Northern European group, British marshes are different in that the seaward edge is limited to mean high tide and spring tide level, and thus not inundated daily. Vegetation structure is different in that *Sarcocornia* spp. also act as pioneers, occupying the seaward edge with *Spartina* spp. (Allen, 2000; Pratolongo et al., 2009). European marshes are also less rich in organic content than those across the Atlantic, consisting of less than 10%, although a decreasing trend is evident from the early stages of their evolution when they were considered organogenic rather than mineralogenic (Allen, 2000).

The saltmarshes of Denmark have revealed such organogenic formation when MSL was at -4 m relative to present approximately 4,000 years BP, (measured from the vertical datum Danish Normal Null (DNN). From this time it continued to accrete to around 2,000 yr BP, reaching a maximum of approximately -1 m DNN when it was transgressed by intertidal mudflats representing a small highstand followed by a minor sea level fall or regression (Bartholdy, 2012). Southern European marshes such as the Venice lagoon are dominated by *Sarcocornia* spp. at the seaward edge although *Spartina* spp. are also present

(Day et al., 1998; Pratolongo et al., 2009). Here a palaeo plain was flooded by rising sea level around 6,000 years BP and subsequent barrier islands have created a vast, shallow, micro tidal lagoon. The marshes are tightly constrained by elevation, especially at the seaward edge (Fagherazzi et al., 2006), although elevations vary across sites with environmental conditions, including tidal range (Allen, 2000; Silvestri et al., 2005).

Marshes of the Mediterranean group are composed of a low marsh dominated by *Sarcocornia* spp, a mid-marsh dominated by *Limonium* and *Frankenia* species, and high marsh occupied by *Juncus* and *Artemisia* species. Slow rates of sea level rise since 7,000 to 6,000 years BP has created coastal wetland environments dominated by deltas which are subsiding (Pratolongo et al., 2009). In south western Spain, meso-tidal estuarine marshes are dominated by *Spartina* sp. and *Sarcocornia sp* and saltmarsh facies have a mean thickness of 0.7 m (Flemming and Bartholoma, 1995; Castillo, et al., 2000). In the Tinto – Odiel estuary facies transitions from intertidal plain, low marsh to supratidal high marsh occurred around 960 yr BP. The stratigraphic record contains almost 2.5 meters of saltmarsh sediments that completes the stratigraphic estuarine infill sequence laid down during sea level stabilisation with low accretion rates of 0.11 cm / year (Ruiz, et al., 1998). Saltmarshes in this wave dominated coastal region occur mostly within estuaries formed during post glacial sea level rise and subsequent stillstand and small sea level regression (Borrego et al., 2009).

The salt marshes of Eastern Asia are predominantly situated in deltaic environments that developed since a relative sea level highstand of around 2 – 4 meters, that has since been stable or falling slightly (Pratolongo et al., 2009). These continue to prograde from sediment delivered from the Yangtze River although decline in sediment supply since completion of the Three Gorges Dam in 2003 threatens this process (Chen and Zhao, 2008). Native marsh seaward edges occur relatively high in the intertidal zone above neap tide high water level and are dominated by *Scirpus* spp. (He et al., 2011; Li et al., 2014).

The Australasian group including Australia and New Zealand are also dominated by *Sarcocornia* spp., although on the east coast of the Australian mainland it occurs in association with mangroves that fringe the seaward edge. The mainland region has experienced a small highstand since around 7,000 years BP that declines in magnitude from north to south along the eastern coastline (Lambeck and Nekada 1990, Lewis et al., 2013). Suitable habitat for salt marsh is restricted by low sediment supply to the coastline, maintaining unfilled estuaries due to the continent’s arid interior compared to many other regions occupied by salt marsh (Pratolongo et al., 2009). In comparison to northern hemisphere locations, Australasia did not experience extensive ice sheet coverage that delivered large volumes of sediment to estuaries during deglaciation. In this region, saltmarsh sea level reconstructions have been limited to identifying modern sea level change with anthropogenically induced climate change (Gehrels et al., 2012).

Indications of the controlling factors peculiar to Tasmanian native saltmarsh evolution can be gained from comparison with the invasive *Spartina anglica* marshes introduced to Tasmania in 1947 (Pringle,1993; Sheehan and Ellison, 2014). The organic content and sediment accretion rates of invasive *S. anglica* saltmarshes were higher, than those of the native saltmarsh in the Mersey Estuary (Beasy and Ellison, 2013), and showed greater lateral development compared to native marshes in the Tamar Estuary (Sheehan and Ellison, 2014). In this estuary, the seaward edge of *S. anglica* marsh occurs below MSL and marsh extent has reached 4.5 km^2, over one tenth of native marsh extent across the entire island (Sheehan and Ellison, 2014). The differences between native and *Spartina* marshes show that native marshes are slower to develop, occupy a narrower elevation range and are likely to be minerogenic, depending on allocthonous sediment supply to maintain their position above tidal flat elevations. These findings are consistent with comparisons between native marsh and invasive *Spartina* in Eastern Asia where the latter demonstrated higher accretion rates (Li et al., 2014), longer annual growth periods, greater above and below ground biomass production, salinity range, inundation period and depth tolerance (He et al., 2011).

Clear differences in the form and function of native Tasmanian salt marshes and those dominated by *Spartina* species are apparent. Furthermore, a consistent relationship between salt marsh species occurrence and a specific tidal datum has not been demonstrated globally. These indications are not consistent with the prevailing view that salt marshes are confined to specific tidal planes or that there vertical extent is directly proportional to tidal range (McKee and Patrick, 1988; Kirwan and Guntenspergen, 2010). Consequently, sea level change studies seeking to use the sedimentary record, should commence with detailed knowledge of its form and function related to mean sea level.

2.7 Conclusion

In tectonically stable Tasmania, the primary cause of Holocene relative sea level change was likely to have been eustatic sea level rise. Paleao-environments of deposition in Tasmanian estuaries need to be located and interpreted within the accepted geochronological framework for south eastern Australian estuarine evolution (Sloss et al., 2004; 2006a, b; 2010). Sea level index points derived from estuarine sites that utilise transgressive sedimentary sequences should be developed from detailed biological and lithological stratigraphic analysis that can be accurately linked to present mean sea level (Sloss et al., 2007). Such an approach will help identify any local variations in the modern landform’s evolutionary pathways, as consequence of sea level change and other local forcing factors and boundary conditions.

Chapter 3 Comparison of biogeomorphological zonation of Tasmanian salt marshes

3.1 Introduction

Saltmarshes have been the key to reconstructing Holocene sea level change across the globe (Redfield, 1972: van de Plassche, 1986; Lambeck, 2002; Horton and Edwards, 2006; Barlow et al., 2014), because sediment deposition and surface patterns of biological zonation are intrinsically linked to the physical and biological processes controlled by tidal inundation period (Adam, 1990; Allen, 2000; Cott et al., 2012). These biogeomorphic processes are recorded in the sedimentary record that salt marshes preserve as faunal, floral and lithological fossils.

Although saltmarsh surface form appears similar the world over in any setting (Bartholdy, 2012) their depth and age reflect the interaction between rates of sea level change and both marine and terrestrial sediment supply, which may all vary over time within the antecedent topography. At the beginning of the Holocene, a warmer global climate and higher rainfall caused rapid erosion and sediment to be transported to low lying coastal locales, filling estuaries and increasing surface elevation within the tidal frame. The marshes that form on these deposits have a dynamic equilibrium with mean sea level position, and the patterns of biological zonation that reflect the relationship are used to identify the spatio-temporal nature of sea level change, from fossil features within the stratigraphic record (Wolanski et al., 2009).

Plant zonation is the most readily observed pattern across saltmarshes but is now rarely used as the feature linked to mean sea level (MSL) for reconstruction purposes. Rather, the distribution across the intertidal gradient of foraminifera is used to build a species and elevation modern analogue that is applied to interpret the fossil record (Gehrels et al., 2000, 2002; 2008; 2012; Horton and Edwards, 2006; Milker et al., 2015). Diatoms are also used (Gehrels et al., 2001; Zong and Horton; 1999) and more recently, testate amoeba have been shown to provide comparable performance values in transfer function applications (Barnett et al., 2016). However, vegetation description usually precedes the analysis of these micro-organisms (Scott and Medioli, 1978; Zong and Horton, 1999; Gehrels, 2002; Engelhart et al., 2011a; Kemp et al., 2012; Barlow et al., 2014), because it guides their sampling and description with visual representation of the elevation zonation concept (Hayward et al., 2010), and then serves to orient the reconstruction results in present time and space (Hartig et al., 2001; Milker et al., 2015). Indeed, for the northern hemisphere, there is over a century of literature on saltmarsh biogeomorphology that has been drawn on to continuously build sea level history (Oliver, 1913, Carey and Oliver 1918 and Cozens-Hardy, 1924 all in Pethick 1980; Steers, 1964; Ranwell, 1964; Redfield 1965; Kestner, 1975; Shennan, 1982; Pethick, 1981; Scott and Medioli,1980; Gehrels et al., 2001, 2002, 2008; Barlow et al., 2015). The present study seeks to elevate our understanding of Tasmanian salt marshes at a regional level, beyond the widely held generalities of marsh form and process, to similarly build a capacity to accurately reconstruct Holocene sea levels.

A regional approach is consistent with evidence that, although detailed investigations of the reconstruction site is essential, distinct differences in biotic responses to physical factors occur between sites (Gehrels, 2002; Silvestri et al., 2005; Woodroffe and Horton, 2005; Saintilan and Rogers, 2009). Furthermore, modern analogues that include a broad range of physical factors can provide more accurate results (Gehrels, 2001; Leorri and Martin, 2009; Watcham et al., 2013; Barlow et al., 2014), because site specific data only captures one point in time. Resolving for the temporal limitation in modern analogues is important because coastal sedimentary landforms and their biota mutually co-adjust form and process with sea level change, including during periods of stillstand (Bird, 2000; Woodroffe, 2002; Sloss et al., 2007; Lewis et al., 2013). Furthermore, site specific studies that do not encompass species-environment interactions with climatic variation have been suggested to limit the ability to predict future environments under climate change (Pennings and Moore, 2000) and this could also be interpreted as a limitation to reconstructing environments under past climates.

A regional approach can also make clearer the generalities around salt marsh form and process that are useful for general conceptual understanding (Pennings and Bertness, 2001), but have implications for sea level studies from regions where little in known, because tidal plane divisions are not made the same everywhere (Eisma, 1997; BoM, 2016). For example, it has been widely held in the literature that mangroves occur between mean tide and high tide elevations, but this correspondence was seldom demonstrated until recently (Ellison, 2009). Similarly, saltmarsh seaward edges are said to occur between mean high tide level, and highest tide level. However, these specific tidal planes are not consistent in the literature. Pratolongo et al., (2009), who adopts the continental scale groupings first presented by Chapman (1960), and includes the geomorphological classifications of Pye and French (1993), shows that tidal hydrology as a key driver of salt marsh form and process is variable at local, regional and global scales.

The US Atlantic east coast marshes are dominated by *Spartina* at the low marsh and seaward edge, at an elevation below mean low water in a macro to meso tidal range, and are inundated daily for up to four hours (Pratolongo et al., 2009). The northern European salt marshes, are also dominated by *Spartina*, but the low marsh is limited to mean high tide to spring high tide levels (Adnitt et al., 2007). Southern European marshes in microtidal settings such as the Venice Lagoon are dominated at the seaward edge by *Sarcocornia* although *Spartina* is also present, (Day et al., 1998; Pratolongo et al., 2009). In these marshes, zonal elevation, hence inundation period, is variable across sites with mean low marsh elevations ranging from 3 cm to 15 cm above MSL, which is around mean high tide level where the tidal range is little more than 1 m (Silvestri et al., 2004). East Asian native salt marshes occur high in the intertidal zone above neap tide high water level and are dominated by *Scirpus* (He et al., 2011; Li et al., 2014). Salt marshes of Australia and New Zealand are dominated by *Sarcocornia* in the low marsh, but the seaward edge is fringed with mangroves up to around high tide level for the Australian mainland (Saintilan 2009; Saintilan et al., 2009).

Presenting the relationship between tidal inundation and zonation by plotting tidal inundation period against elevation and marking on it a species upper and lower limits, enables comparison across sites, particularly tidal ranges (Ranwell et al., 1964). Later Adam (1990) used the sigmoidal inundation period curve to indicate ‘critical tide levels’ for certain biota. However, the application of the model was not confirmed by some later studies (Pennings and Bertness, 2001). While few vegetation studies express their results in in terms of tidal inundation period, Gehrels et al., (2001) show that for sea level studies using foraminifera and diatoms, zonation does conform to tidal inundation period across different tidal ranges. As an unambiguous, continuous variable, tidal inundation period should be re-considered for regional comparison of biogeomorphological features, rather than the arbitrarily defined tidal plane categorical data.

There are three key biogeomorphological features related to MSL that can be identified as facies transitions within the sedimentary record that are controlled by patterns of inundation period peculiar to each site: the seaward edge of marsh, transitions between low and high marsh, and the transition from marsh to freshwater wetland or terrestrial vegetation.

3.1.1 Transition from tidal flat to saltmarsh

The transition from tidal flat to saltmarsh seaward edge contains important modern analogue information on process threshold conditions related to MSL (Pethick, 1980; Thomas and Varekamp, 1991; Allen, 2000; Woodroffe, 2002; Sheehan and Ellison, 2014; D’Alpaos and Marani, 2015). Fagherazzi (et al., 2006), models the mechanisms responsible for the initiation of salt marsh from tidal flat, constraining the relationship between water depth and shear stress that is required for their development and maintenance respectively, see Figure 3-1. Between tidal flat and saltmarsh, a zone of semi stable equilibrium exists that occupies a narrow elevation range. Tidal flats are described as dynamically stable, because in deeper water, higher waves increase bottom shear stress that causes erosion. Erosion continues until a depth where wave oscillation effects do not reach the bottom, and deposition is initiated and persists until depth is reduced to a threshold, where oscillation effects again induce erosion.

Shallower water tidal flat elevation is controlled by sediment deposition from suspension, because at these depths bottom shear stress is limited by wave dissipation from bottom friction, depth induced breaking and white-capping. Here the tidal flat exists in unstable equilibrium because net deposition will continue until SM develops. The region of continuous shear stress is then the zone between deeper water tidal flat and saltmarsh, and is thus dynamically stable. The model notes the difference in shear stress according to sediment type, and this could cause variation in the elevation distribution of the three intertidal zones across different geomorphic settings. The model is also applicable to the processes involved in marsh progradation by seaward edge plant pioneers.

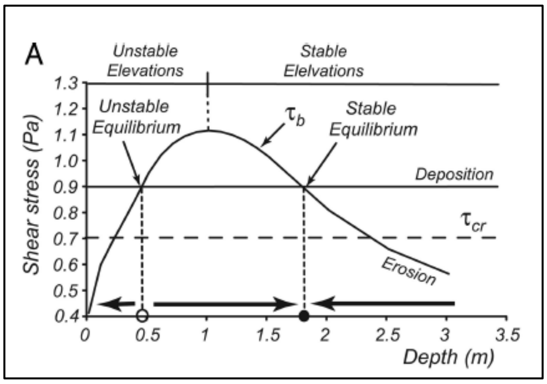


Figure 3-1 The distribution of bottom shear stresses with depth showing that it peaks at intermediate elevations, and is a point of semi stable equilibrium Stable equilibrium occurs in the lower tidal flats where deposition and erosion are balanced thus elevations are maintained. At depths shallower than the intermediate zone elevations are in unstable equilibrium because as shear stress declines, deposition is greater than erosion and the tidal flat continue to accrete until salt marsh develops. **Source: Adapted from Fagherazzi et al., (2006).**

In the saltmarsh, numerous evolutionary pathways beyond the threshold phase occur, related to sediment supply and type (Haslett et al., 2003; Fagherazzi et al., 2006), antecedent condition and accommodation space (Schwimmer, 2001; Cott et al., 2012), and rates and direction of sea level change or stability (Schwimmer and Pizzuto, 2000; Wolanski et al., 2009). Plant species life form affects sediment baffling capacity which affects sedimentation rates (Ranwell,1972; Adam,1990; Konig, 1948 in Langlois *et al*., 2003), while accretion rates are affected by differential below ground biomass accumulation Cahoon et al., 1999; Allen, 2000; 2002; Rogers and Saintilan, 2009; Wolanski, et al., 2009; D' Alpaos and Marani, 2015).

3.1.2 Transitions between zones within the marsh

Transitions between zones within the marsh are also used to indicate change in the position of the tidal frame (Gehrels, 2002; Haslett et al., 2003) because abiotic factors associated with tidal inundation contribute to complex environmental gradients of soil salinity and drainage (Snow and Vince, 1984; Sanchez et al., 1996; Haslett et al., 2003; Wolters et al., 2008; Saintilan, 2009; He et al., 2011; Davy et al., 2006; Mahall and Park, 1976; Day et al., 1988; Castillo et al., 2005). Soil salinity and drainage have been shown to be affected by the texture of sediments (Maricle and Lee, 2007). Whilst tides exert the primary long – term control on soil salinity and soil moisture (Mahall and Park, 1976), field studies are limited to capturing snapshots of conditions and limited evidence of a direct relationship between soil salinity and inundation period has been shown even with seasonal sampling (Ranwell et al., 1964). Regional scale comparisons have indicated climate effects on species dominance and zonation, due to differential evaporation and precipitation (Pennings et al., 2003; Ewenchuck and Bertness, 2004). Other abiotic factors include soil redox potential and nutrient content (Judd and Lonard, 2002).

Strong biotic controls, such as competition between species, also affect zonation (Bockelman and Neuhaus, 1999; Hacker and Bertness, 1999; Dormann et al., 2000; Rand, 2000), although relative competition intensity can be influenced by physical factors related to elevation (Huckle et al., 2000). The lower limits of salt marsh species are set by tolerance to the physical conditions controlled by elevation, whilst the upper limits are set by biotic interactions controlled by competitive relationships between species (Pennings and Callaway, 1992). In support of this generalisation, Dormann et al. (2000), identify an interaction between competition, herbivory and zonation and Hacker and Bertness (1999), show that a mid-marsh community requires less harsh conditions than the lower elevation zone, the absence of competitors and the presence of a facilitator species that ameliorate soil conditions, which would otherwise be limiting. Engels et al. (2011) show that competition between glycophytes and halophytes occur with early seedling establishment. Seedling success is related to relative levels of salinity and inundation period, with a strong interaction between these two factors. Competition at the seedling and recruitment stage could well explain the sharp boundaries between most zones (Rand, 2000). However, consistent associations between boundaries and soils have not been found (Pennings and Callaway, 1992; Silvestri et al., 2004).

The physical and biological interactions described above are recorded in the sedimentary record because the biogeomorphic responses to inundation period with sea level change involve internal feedbacks that maintain elevation of the marsh relative to the tidal frame (Redfield, 1972; van de Plassche, 1986) and the areal extent of marsh (Schwimmer, 2001; Schwimmer and Pizzuto, 2000), as a result of either site specific allocthonous or autocthonous surface processes (Cahoon et al., 1999; Rogers and Saintilan, 2008 & 2009; Lovelock et al., 2011; Cott et al., 2012). Clearly, inundation period is the primary variable controlling these processes, but the patterns of biogeomorphic features used to indicate MSL are also the result of the manner in which it interacts with other parts of the morphodynamic system. Pennings and Bertness (2001), state that determining whether the patterns of variation in inundation period globally is a cause of the variation in patterns of biological variation is worthy of further investigation, a question that remains uninvestigated.

Understanding this variability so as to quantify the relative role of controlling factors is needed for high precision local sea level change reconstructions. These reconstructions are vital for predicting future sea level change and its likely impact on coasts (Allen, 2000; Rogers and Saintilan, 2009; Wolanski, 2009; Cott et al., 2012; Rogers et al., 2012). In Tasmania, very little is known about sea level history. Pennings et al., (2003) state that, for ecological studies to be reliably predictive, they need to be conducted at a broad scale, across regions, so that evolutionary processes, such as adaptation of species to local factors can be perceived. This is just as important for palaeoecological studies, because if geomorphic evolutionary processes are used to reconstruct past sea levels, then it is important to understand what variation might mean in terms of error when inferring processes from the sedimentary record, or interpreting missing analogues.

The present study aims to determine the relative influences of tidal inundation period, sediment type and climatic variation across regions, on salt marsh biogeomorphic zonation in Tasmania. The objective is to understand if biogeomorphic transition points can be related to certain tidal planes and contain features that may be distinguished in the sedimentary record as facies transitions to infer past sea level position. To achieve this objective, saltmarshes with different tidal ranges, geomorphic settings and therefore sediment type, and climatic conditions were compared and contrasted.

3.1.3 Zonation in Tasmania

Tasmanian salt marshes provide an excellent location to test generalisations, because, in comparison to the lower diversity marshes of the Northern Hemisphere, particularly those composed of nearly monospecific stands of *Spartina*, Tasmanian marshes contain up to 5 distinct vegetation zones. Little is known regarding the interactions between the topographic, hydrologic and edaphic factors limiting them, nor the degree to which their elevation ranges vary as a consequence of these interactions (Saintilan, 2009). Therefore, generalised response models (including those for palaeoreconstruction) to sea level change, based on elevation, contain considerable sources of uncertainty. Adding to the uncertainty, is that of Tasmanian marshes range from micro – to small macro tidal, and exist in a range of settings from the upper reaches of estuaries to low relief open coast settings. Saintilan (2009) describes that the Australasian patterns of salt marsh species richness at a continental scale do not conform to biogeography theory (Taylor and Gaines, 1999; Whittaker et al., 2001) in that species richness increases with decreasing intertidal area, with higher latitude and declining temperature (Saintilan, 2009).

For some zonal dominants, eco-physiological responses of closely related vicariate species are known from similar cool temperate settings (Chapman, 1960; Adam, 1990). In the present study, the known responses are used to help determine the environmental variables that are likely to control zonation in Tasmania, including seaward and landward limits and to help select model variables.

3.1.4 Study sites

Three study sites were selected in settings with unrestricted, permanent access to the ocean (Figures 3-2b, c and d). After reconnaissance of saltmarshes around the island, study sites were selected that displayed a low degree of human disturbance such as, urbanisation, upstream water diversion and storage for agriculture or livestock grazing. The three study sites are situated within small catchments in which shell fish and livestock grazing industries share the natural resources according to the regulations set out in statutory Water Management Plans (DPIPWE, 2004). The plans implement rules for freshwater use that does not compromise the quality or quantity of flows to the estuary (DPIPWE, 2004). At each site the saltmarshes displayed geomorphologically mature features such as a well-defined tidal creek rather than numerous small ephemeral types and clear, consistent zonation. Sites were also checked for the depth of salt marsh sediment to evaluate their maturity.

Site one in Luttrells Bay and site two in Pelican Bay are located on the south eastern coast of Tasmania, and have micro tidal ranges and similar climate. They differ in salinity because of differences in their proximity to marine influence, lagoon depth, wave energy and sediment type. Site three, Boullanger Bay, is located in far north western Tasmania and has a meso tidal range and oceanic salinity (Table 3.1 and Figure 3-2b, c and d). It is cooler than the east coast sites with higher annual and growing season rainfall. It has the same dominant vegetation types as the eastern marshes, but with greater species diversity due to heterogeneous micro-topography, reflecting its more open coast setting and higher energy conditions (Table 3.1).

Luttrells Bay within the Little Swanport Estuary is set between the fluvial and bay head deltas of the upper, muddy reaches of an open drowned river valley estuary. Fresh water flow into the estuary is intermittent and only occurs during high rainfall events. Thus, terrestrial sediment supply is also limited. Pelican Bay is a shallow embayment situated close to the mouth of an open barrier estuary. Its larger catchment is not heavily dammed, thus providing a continuous fluvial sediment supply.

Table 3-1 Tidal plane elevations relative to mean sea level and the physical factors likely to influence salinity, pH, sediment type and moisture content at the three study sites. Note that the tidal plane terms are different at Boullanger Bay because of the stronger diurnal variation.

0

Feature	**Site 1 Luttrells Bay**	**Site 2 Pelican Bay**	**Site 3 Boullanger Bay**
Latitude, Longitude	-42.3499, 147960191:866,686	-42.0749, 148.21751:866,686	40.73847, 144.849591:27,084
Geomorphic setting	Upper reaches of drowned river valley estuary	Seaward end of barrier estuary	Open sandy coast
Catchment size	600 km^2	900 km^2	
Likely tide salinity	Brackish	Higher salinity due to shallow lagoon and close proximity to marine source	Oceanic salinity
Reported Tidal range (m) HAT MHHW MLHW MLLW LAT	1.2 0.754 0.605 0.029 -0.605 -0.827	1.2 0.733 0.615 0.035 -0.615 -0.84	3.76 1.756 3.174 (MHWS) 0.996 -1.242 (MLWS) -1.932
Average annual rainfall & minimum rainfall (mm)	559 – 273	688 - 381	948 - 617
Average high temperature (Jan)	24.5C	24.6°C	22.7°C
Surrounding topography	Low valley slopes	Flood plain	Coastal plain

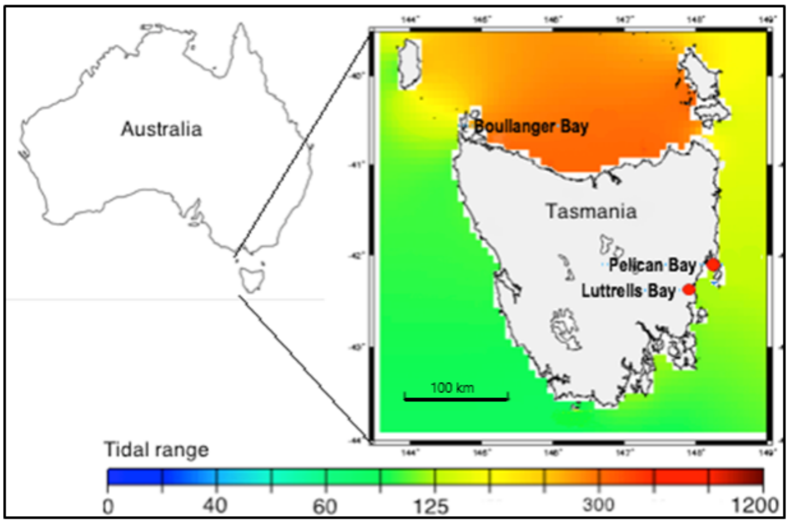
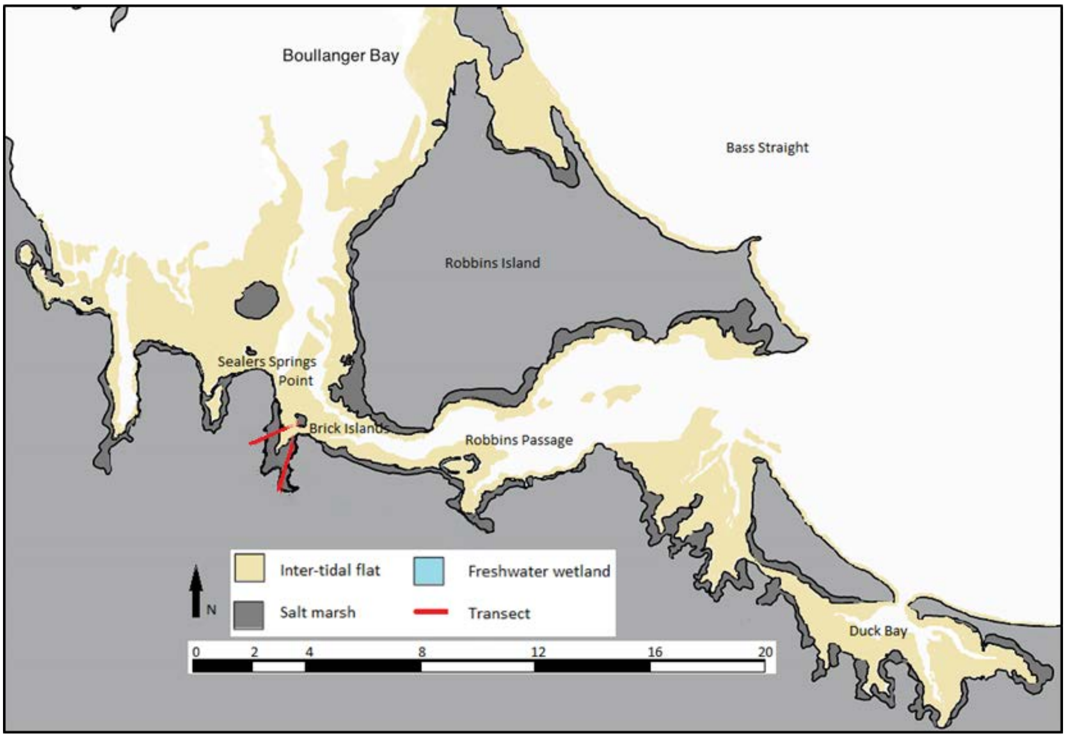


Figure 3-2 a) Location of study sites at Boullanger Bay, Luttrells Bay and Pelican Bay in Tasmania showing the tidal range around the island. Source: BoM National Tide Centre 2013



3-2 b) Transect locations at Sealers Springs Point, Brick Island and Robbins Passage within the Boullanger Bay open coast site.

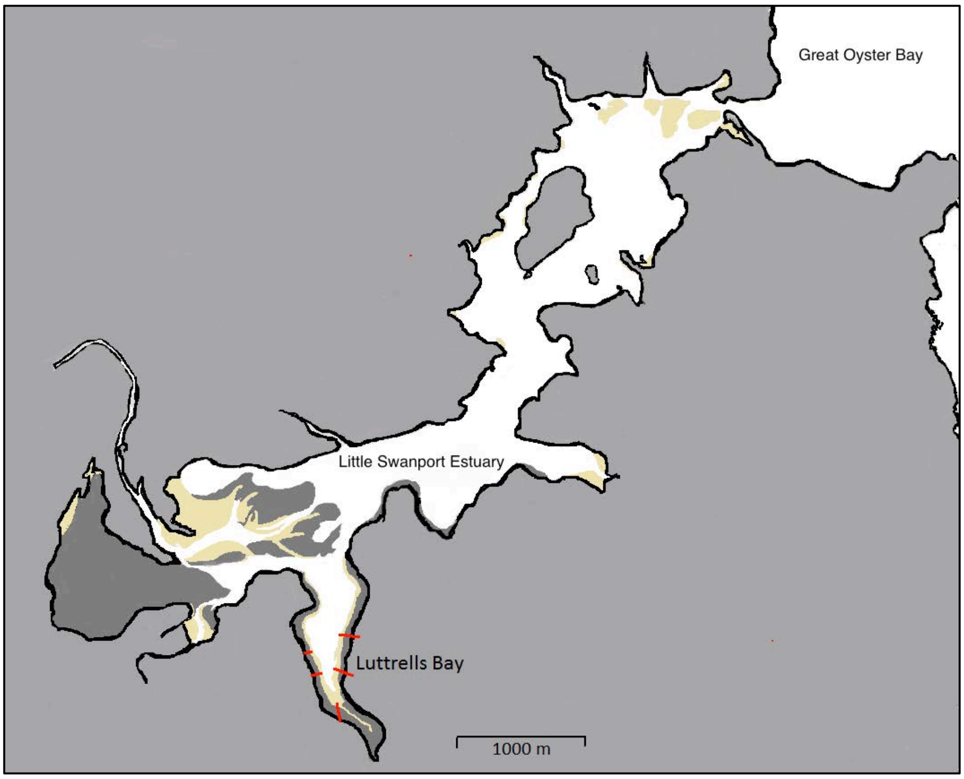


Figure 3-2c Luttrells Bay study site and transect locations within the Little Swanport Estuary drowned river valley.

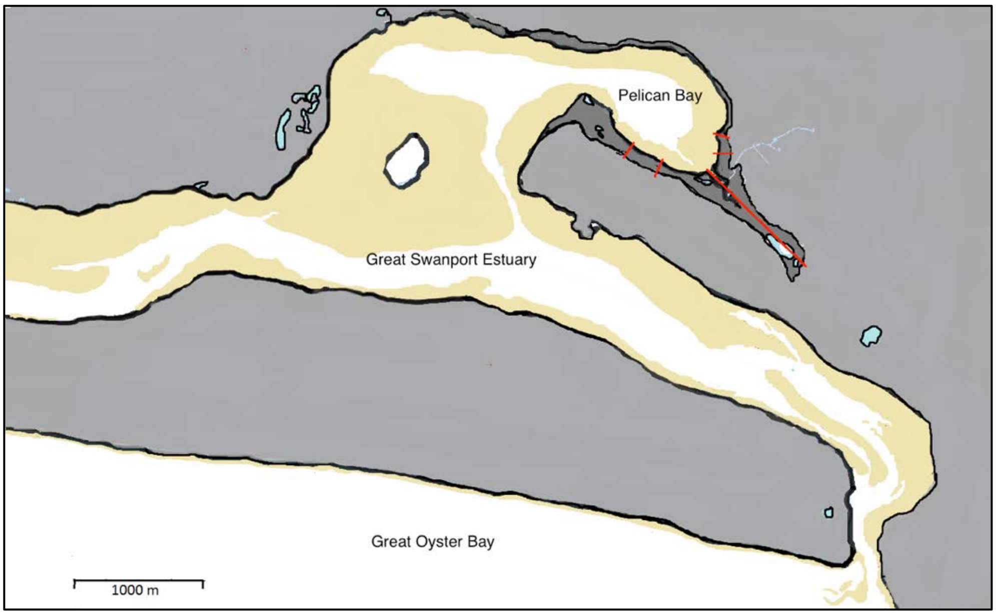


Figure 3-2d Pelican Bay study site and transect locations within the Great Swanport Estuary, a back-barrier lagoon site.

3.2 Methods

3.2.1 Modelling biogeomorphic zonation

Modelling the causality of salt marsh biogeomorphology is difficult because zonation is controlled by closely coupled environmental variables that can behave co-linearly with the elevation gradient (Davy et al., 2011). Co-linearity violates one of the fundamental assumptions in predictive modelling with multivariate data (Quinn and Keough, 2002), so excludes many of the typical predictive regression models, mainly because statistical analysis cannot separate causality when two or more independent variables are highly correlated. However, with recent advances made in understanding the eco-physiological responses of some zonal dominants (Rand, 1976; Davy, 2006; Naidoo and Kift, 2006) described as vicariate across continents (Chapman, 1960; Allen, 1990), this can be overcome by only using the most likely variables, or surrogates, for a particular species that are not highly co-linear. This parsimonious approach can provide numerically feasible descriptions that retain the most relevant features of the governing processes.

For the analysis, as much environmental data as practicable was collected along the elevation gradient including soil moisture, salinity, pH and sand, silt and clay content and was checked for co-linearity. Elevation was converted to inundation period after Ranwell (1964). Scatter bi-plots showed percent sand, silt and clay were highly co-linear with inundation period, soil moisture and salinity, so were removed from the data set. Notwithstanding that tidally deposited sediment becomes an interactive element in the morphodynamic system by way of influencing elevation, (Viles, 1988; Cahoon et al., 1999; Bird, 2000; Woodroffe, 2002; Silvestri, 2005) removing sediment texture from the analysis was considered reasonable because, rather than being a cause of biological zonation, sediment variation in small tidal ranges is likely to be a consequence of salt marsh biogeomorphic processes related to tidal inundation period. Such processes include both the physical transport and deposition of sediment by tides and biogenic sediment production, the rate of which is influenced by vegetation type (Beasy and Ellison, 2013; Sheehan and Ellison 2014).

Five distinct vegetation zones were observed in the field, and the change in species dominance between them occupied a narrow geographic area and is termed hereafter the transition point. The a priori classification of the five vegetation zones was first tested.

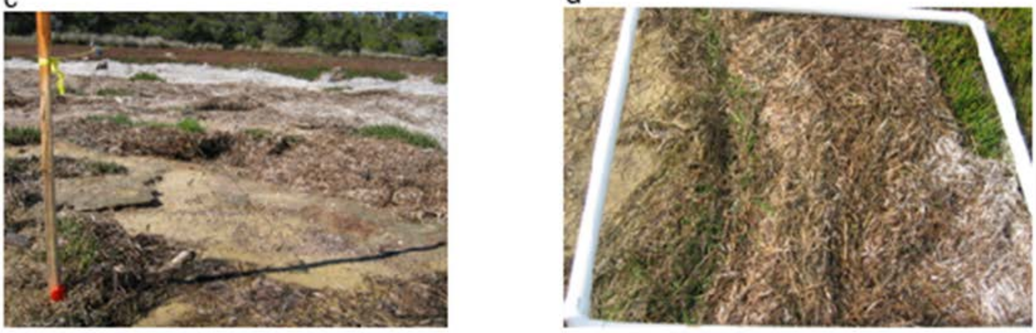
3.2.2 Transects, elevations and inundation period

Aerial photographs of each site were examined, along with prevailing wind and wave direction information, to capture geomorphic variability within sites. At all sites the strongest and most persistent winds are north westerlies. To position transects, marshes were divided into 2 physiognomic units, to encompass the leeward and windward side of the marsh and transects positioned orthogonal to the coast, (see Figures 3-3 a to f). Survey control points were installed at each site, using three differential GPS base stations to tie in transect elevation profiles to mean sea level, according to the Australian Height Datum.

a b

Figure 3-3 Seaward edges at Luttrells Bay a) leeward side of marsh showing micro cliffed cleft and neck formation, not inundated daily. B) windward side showing slumping edge and inundated daily
both dominated by *S. repens*

c d



At Pelican Bay c) windward side of marsh showing eroding edge by burial with wrack, inundated daily and dominated by *S. quinqueflora*. **d)** leeward side of marsh with chenier type edge, dominated by *S. quinqueflora*, not inundated daily

e f

At Boullanger Bay e) partly sheltered site showing prograding marsh edge dominated by *S. quninqueflora*. **f)** open coast site showing receding marsh edge dominated by *S. quinqueflora*.

Sampling was conducted to achieve two objectives; the first being to describe the edaphic features of each zonal transition point to support interpretation of the predictive analysis in this study. The data set is termed hereafter the transition point data set. A second set of samples was collected to provide data for the predictive analyses and termed hereafter the multiple logistic regression (MLR) data set. To populate both, five belt transects each four meters wide were deployed at each site, perpendicular to the intertidal gradient and surveyed for elevation relative to MSL. For the transition point data set, four of the five transects were surveyed using a Topcon Electronic Total Station, GTS 603. Each transition point for a zone was recorded as the first point where the dominant species occurred on each side of the belt transect, (see Figure 3-3). The elevation gradient of the fifth transect as each site was surveyed with a Leica NA270 Level for the MLR data set, to locate every 2 cm vertical rise in elevation for sediment sampling. Two-centimetre vertical increment sampling is the widely accepted standard for a biological training set, because it is likely to capture change across geomorphic zones, related to tidal inundation period (Gehrels, 2008).

Inundation period for all sampling points was calculated as the percentage of time that sea level exceeded each elevation, using six minute sea level observations from the nearest tide gauge location from 1992 to 2011 (data provided by the Australian Bureau of Meteorology). The modelled difference in tidal range from lowest astronomical tide to highest astronomical tide from the Burnie tide gauge to Boullanger Bay was – 0.11 m (BoM, 2016) with no expected tidal attenuation due to the open coast setting. The difference between the tide gauge at Spring Bay and Little Swanport and Moulting Lagoon estuary mouth were + 0.04 m and 0.06 m respectively (BoM, 2016). Anecdotal evidence from oyster growers at the southern sites indicated an indiscernible difference in the tidal range from the Moulting Lagoon estuary mouth to Pelican Bay, and a less than 0.1 m variation in tidal range between the Little Swanport Estuary mouth and Luttrells Bay (H. Dyke, 2015, pers.comm). The potential error range for inundation period calculations is greatest at Luttrells Bay but not exceeding 0.14 m across the 1.2 meter tidal range. However, field observations indicated the modelled inundation periods to be consistent with reality.

3.2.3 Vegetation data collection

Vegetation abundance data were collected across the marshes within 1 m^2 quadrats (determined by the species area curve techniques), for all vascular plant species present at all sites. These were estimated as a percentage of the quadrat occupied by the above ground part of each species using the Domin 10 point cover scale (Kent and Coker, 1992). Field observation showed five distinct vegetation zones across three marsh divisions, low, mid and high marsh. Each zone was comprised of one to two dominant species in the tallest stratum. Zones 1 and 2 comprised a succulent herb-field low marsh, zone 3 a sedge-land mid marsh, and a grassy high marsh was discriminated from dominant life-forms in zones 4 and 5. Species accumulation curve analysis identified that 22 replicates were required to adequately represent each zone, so 25 replicates for each zone were taken to account for differences in area between sites. This was repeated in each physiognomic unit. Sampling points were located using random numbers where the first digit dictated the number of meters to the next location. The second digit determined the number of meters in from the seaward zonal boundary of the vegetation zone. A total of 250 replicates were sampled at Luttrells Bay and Pelican Bay, and 175 replicates at Boullanger Bay due to narrower zone depths and reduced access to the remote site during daily high tides.

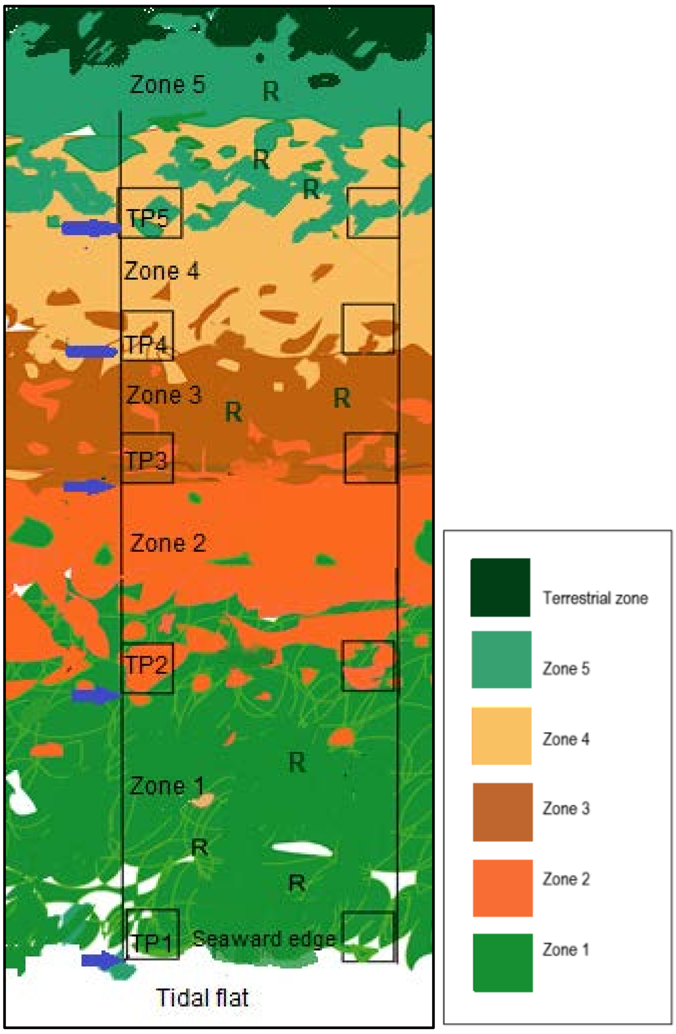


Figure 3-4 Representation of the belt transect sampling design, from the tidal flat to the terrestrial zone (adapted from Morrison, 2006). Squares represent the 1 m^2 quadrat placed on the first point at which the dominant vegetation of that zone occurred on both sides of the belt transect. R represents the randomly located vegetation and sediment sampling points for the MLR data set, and blue arrows show the location of the transition point sampling sites. The total replicates were mostly retrieved from outside of the transects and spread across the whole marsh area.

3.2.4 Sediment chemistry and texture

Because the dominant genus of the low and mid marsh at all sites, *Sarcocornia* is dormant due to frost sensitivity through late autumn, winter and early spring, seasonal sampling was not neccessary. Mahall and Park, (1976) found little seasonal variation in soil moisture and salinity across zones in similar settings in Spain. Sampling at all sites was conducted in January to February, to maximise differences in salinity, pH and moisture content between vegetation zones.

Sediment texture and chemistry analyses for the transition point and MLR data sets differed for two reasons. A high level of precision was required for the smaller transition point data set to reduce noise, and because different sediment types were expected across the intertidal gradient. Secondly, the larger MLR data set meant that sediment sample sizes had to be smaller due to sediment collection permit requirements, so the methods were adjusted. For the transition point samples, the soil saturation paste / solution extraction method offered precision because it accommodates all sediment textures. Distilled water is added to dried sediment until saturation is indicated by water pooling on the sediment surface (Sparks et al., 1996). The salt content is calculated as a proportion of the amount of solution extracted from surface sediment samples weighing 150 grams. Four replicate samples from each transition point along transects were retrieved at each site (n=60), and sieved through 200 micron mesh to remove plant material, shells and clasts. Sediment texture analysis at the same location was conducted in the field using the texture by feel method (n=60), after Thein (1979).

For the MLR data set, 60 gram surface samples (5 cm depth) were taken at each 2 cm vertical increment along the 5th transect at each site (n=112). Moisture content was measured from 5 ml of sample that were weighed, dried at 60° C for 48 hours and reweighed for moisture content. Samples were then pulverised and added to 40 ml of distilled water, shaken and left for 24 hours, and a further 200 ml was added before measuring salinity using a EZDO 7200 Multiparameter Probe. For grain size analysis, the hydrometer method was employed to quantify the relative proportion of sand, silt and clay after Gee and Bauder (1986) using 40 gram samples. For pH, dampened pH paper test strips (with 0.2 pH increments) were used and pressed into another 5 ml of sediment within 24 hours of collection. Three replicates were taken and averaged. Comparison with a pilot run of pH in solution, showed pH test strips provided less variable and more realistic results on the small samples than the pH in solution method.

3.2.5 Data handling

The primary aim of the statistical analysis was to understand whether marsh divisions, determined through vegetation analysis and description, could be predicted with measurable controlling variables, and to tease out the relative influence of each variable at a site and regional level. A second aim was, to establish if the physical traits at the transition point between these zones were different, such that their interpretation as facies contacts within the sedimentary record might be possible.

For the vegetation data set, cluster analysis using Ward’s Method and Bray-Curtis distance on the abundance codes of species was used to determine biogeomorphic zones. A dendrogram was produced to visually confirm the number of groups, and Analysis of Similarities (ANOSIM) was used to test for compositional differences between a priori and numerically classified zones.

For the transition point data set, Multiple Analysis of Variance (MANOVA) was employed to test for a significant difference between soil pH and salinity taken together, between zone transition points, across all three sites. MANOVA was selected because pH and salinity were measured from the same solution extracted from each sediment sample, and so are not independent of one another (Quinn and Keough, 2002). Furthermore, the analysis controls for the probable but moderate correlation between the two (Mertler and Reinhart, 2017). The analysis compares the differences between groups from the vectors of the two means, by creating a new dependent variable from them, which maximises the differences between groups (Mertler and Reinhart, 2017). Data were transformed to improve normality, but outliers were retained to enable comparison of sites which were considered to have ecologically important information. Pillai’s Trace test statistic was employed to test the significance of differences between groups because it is robust when the sample sizes are small and there is unequal variance and covariance across groups (Quinn and Keough, 2002).

Multiple Logistic Regression (MLR) was employed to predict the presence of the high, low and mid marsh at each site and then across sites. The MLR is a generalised linear model that is widely used in a predictive capacity for ecological studies (Trexler and

Travis, 1993). It provides a flexible framework for the analysis of binary or dichotomous data (Juggins, 2007) that ordinary least squares regression cannot model, and is able to rank predictors in order of the degree of influence they exert on the response variable. It accommodates non-normal distributions of response variables and error terms, because logistic regression estimates the likelihood that the model is most like the data, rather than minimising the distance between the model and the error contained in the data, thereby requiring normality of both. This includes data that models best when not transformed to meet normality assumptions, (Quinn and Keough, 2002) as was the case in this study. Furthermore, linearity is not required between the response and independent variables. However, multiple logistic regressions cannot handle near perfect or high multi-co-linearity, for either linear or nonlinear relationships, between independent variables. With co-linearity, interpretation of the relative importance of the correlated independent variable can be unreliable. Causality is most likely to be attributed to the variable which is most accurately measured (Garson, 2008).

Binomial logistic regression was employed to model the relative importance of environmental variables. Stepwise logistic regression was used to remove variables that did not add statistical significance to the model. Although some authors warn against its use (Garson, 2008), Quinn and Keogh (2002) considers the stepwise procedure suitable when applied to pure prediction, rather than theory generation, as is the case in the present study. The logit link function was used to suit the dichotomous response as it links the expected value of y to the predictors...,

$$\delta(\mu) = \beta_0 + \beta_1 X_1 + \beta_2 + X_2 + \ ....$$

$\delta(\mu)$ is the link function and β_0, β_1, etc., are the parameters to be estimated, (Quinn and Keogh, 2002)

Another benefit of logistic regression is that it does not require a balanced design, in that the number of opposing outcomes does not need to be equal. In fact, the analysis works best when the relative proportion of observations is proportional to that in reality

(Agresti, 2002). This was the case with the elevation gradient sampling across marsh divisions with different elevation ranges.

The logistic regression β coefficients were estimated using maximum likelihood (ML) methods (Quinn and Keough 2002) and the statistical significance for each was tested by the Wald statistic. The overall significance of the model was tested by the likelihood ratio test for analysis of deviance, as this accommodates relatively small data sets (Agresti, 2002; Quinn and Keogh, 2002; Garson, 2008). The Cessie and Van Houwelinge (1991) goodness of fit test was employed to evaluate the appropriateness of the model (Harrell, 2001) by testing the significance or otherwise of the unexplained variance in the dependant variable (Garson, 2008).

3.3 Results

3.3.1 Vegetation description and analysis

Thirty-three species were identified across the three sites. Cluster analysis showed five distinct groups that conformed to the *a priori* determination of vegetation zones. Analysis of similarities showed that vegetation zones were significantly different, $R = 0.28$, $p < 0.001$. The dominant species in each marsh zone, which determined the zone's vegetation type, occurred with 80% or greater frequency in zones 1, 2 and 3. The frequency of the dominant species in zones 4 and 5 was more variable, rarely exceeding 80% except in zone 5 at Boullanger Bay (Table 3-2).

Table 3-2 Percent frequency of species at Boullanger Bay (n=125), Luttrells Bay (200) and Pelican Bay (n=175). Species are grouped into the vegetation zone and type in which it was most frequent.

Species	**% Frequency**														
Biogeomorphic zone	**ZONE 1**			**ZONE 2**			**ZONE 3**			**Zone 4**			**Zone 5**		
Vegetation type	LB	PB	BB	LB	PB	BB	LB	PB	BB	LP	PB	BB	LP	PB	LP
Low marsh															
Zone 1															
Sarcocornia quinqueflora low heath															
S. quinqueflora	60	100	100	10	20	20									
***Samolus repens* low herbland**															
S. repens	80		32	85	35	75	25	50	22					15	13
Tecticornia arbuscula	4		7			30									
Zone 2															
***Sarcocornia blackiana* low heath**															
S. blackiana				90	85	91	95	95	78	5	20				
Triglochin striatum				20	10	15									
Hemichroa pentandra			20	6	75	16		40	20	5					
Suaeda australis		12	60	15		58		20			40				
Atriplex paludosa						30			20			22			12
Mid Marsh															
Zone 3															
***Juncus kraussii* sedge land**															
J. kraussii			20	8		46	85	95	89	100	55	81		30	25
Selliera radicans					4		15	70	44	42	60	18		40	
Distichlis distichophylla							60	65	55	20	40	36		40	10
Gahnia filum							20	20							

Rhagodia candolleana							10	5	44	5					10
High marsh															
Zone 4															
***Poa labillardierei* tussock grassland**															
P. labillardierei										75	20	63			5
Lobelia alata								20		60	60	30		60	
Austrostipa stipoides						4	20	5	33	20	25		60	35	
Euchiton spp.										10					
***Leptocarpus brownii* sedgeland**															
L. brownii								40		15	80	15	20	75	15
Isolepis nodosa								5		15	75			65	
Poaceae spp.										40	30	10		60	25
Apium prostratum						16				15	20	20			20
Brachyscome sp.									5		10				
Einadia nutans									5						
Melaleuca ericifolia swampland															
M. ericifolia													10		100
Zone 5															
Terrestrial transition zone															
Lomandra longifolia											25			35	
Pteridium esculentum													35		
Leptospermum sp.										10				25	
Allocasuarina sp.											5				
Acacia sp.										10					
Banksia marginata											5			10	
Villarsia sp.										10					
Eucalyptus spp.										15					

3.3.2 Biogeomorphic zonation: Elevation and vegetation

The leeward seaward edge at Luttrells Bay, transition point 1(TP1) began at a mean elevation of 12 cm ± 2 cm above MSL. The seaward edge exhibited slumping. *S. quinqueflora* and *S. repens* were present in a mosaic distribution and exhibited low plant vigour along the entire leeward edge. The windward marsh edge exhibited micro-cliffing and began at a mean elevation of 18 cm ± 5 cm above MSL, and was dominated by a robust fringe of *S. repens* (see Figure 3-2 a and b). When elevations for the seaward edge types were combined the mean elevation was + 14 cm ± 2.3 cm. At Pelican Bay (PB) and Boullanger Bay (BB) there was no evident systematic change in seaward edge morphology with wind direction. Edges at both sites were dominated by *S. quinqueflora* with a mean elevation of + 39 cm ± 2.2 cm at Pelican Bay and + 86 cm ± 9.3 cm above MSL at Boullanger Bay.

Zone 2 at all sites was dominated by *Sarcocornia blackiana* with less frequent species common to all sites providing varying degrees of species richness. The mean elevation of transition point 2 at each site was 87 ± 2.4 cm at BB (only slightly higher than TP1 at BB due to a chenier type morphology), 37 ± 5.8 cm at LB, and 59 ± 3.1 cm at PB. The most frequent subdominant was *Hemichroa pentandra* at PB, *Samolus repens* at LB and BB contained both in mosaic distributions, and included a high frequency of *Tecticornia arbuscula* (Table 3-2).

Biogeomorphic zone 3 was dominated by *Juncus kraussii* at all 3 sites, and TP3 occurred at mean elevations of 107 ± 9.2 cm at BB, 62 ± 3.5 cm at LB, and 67 ± 4.5 cm at PB. At all sites Zone 4 was dominated by *Stipa stipoides* and *Poa labillardiarei*. TP4 was 132 ± 3.1 cm at BB, 75.8 ± 4.1 cm at LB and 75.5 ± 1.8 cm at PB. Biogeomorphic zone 5 was indicated by the presence of juveniles of the surrounding terrestrial vegetation. At LB this consisted of mixed *Eucalyptus* and *Acacia* species, PB was dominated by *Leptospermum*, and BB by *Melaleuca ericifolia* swamp forest. Figure 3.5 shows the mean elevation of each transition point with SE bars. The box plots show that the elevation of TP 3 and TP 4 are most similar within and between LB and PB. The greatest variation in elevation between the two marshes occurs at TP 1 and TP 2 and these are dominated by different species, hence vegetation types. Notably TP 1 at LB, dominated by *S. repens* is at a much lower elevation than TP1 at PB, dominated by *S. quinqueflora*. TP 2 at LB and TP 1 at PB occur at around the same elevations and are dominated by *Sarcocornia* spp. The plot also shows that the elevation ranges from the seaward edge to the marsh upland border are similar, despite BB having a tidal range more than twice that of the two micro tidal sites.

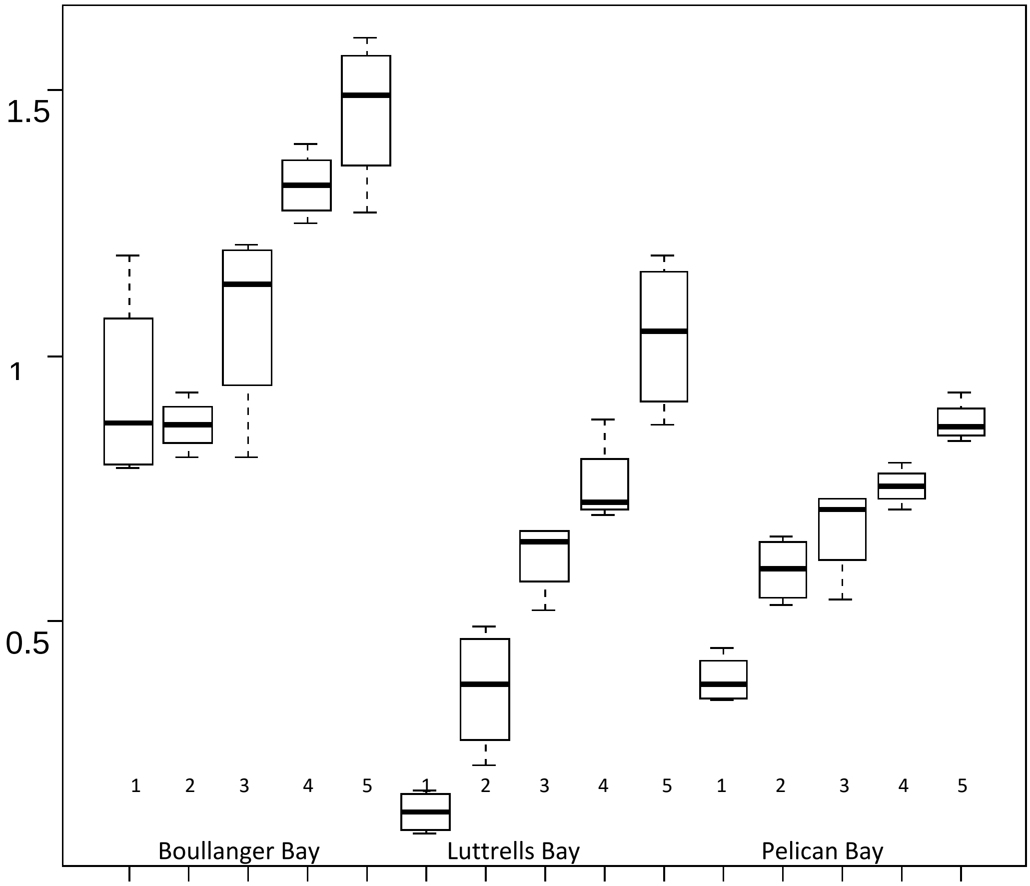


Figure 3-5 Mean elevation with SE (m) of transition points 1 to 5 at each site on the x axis and elevation relative to MSL on the y axis.

3.3.3 Biogeomorphic zone transition point inundation period

When biogeomorphic zonation was plotted against inundation period, similarities between the three sites were apparent (see Figure 3-6). At transition point 5 (TP5), inundation period was less than 0.01 % or 9 minutes at LB and PB, and 0.02 % or 1.7 hours at BB. These transition points are only inundated during meteorologically forced events, such as floods, and extremely low pressure systems in the south (Dyke H., pers comm 2007), and storm surges in the north (author's observations).

The modelled inundation period for TP 4 was 1.3% or 113 hours at BB, 0.12% or 10 hours at LB and 0.04% or 1.2 hours at PB, and field observations supported the variation across sites. This transition point, showed the greatest difference in inundation period between sites. At transition point 3, LB inundation period was 0.6 % or around 52 hours per year and 0.41% or 36 hours at PB. At BB inundation period was 2.8% or around 245 hours per year. Field observations were consistent with this datum.

Seaward from zone 3, at both sites inundation period increases rapidly with decreased elevation relative to mean sea level. Inundation period for TP2 was 8% or 700 hours, and 13 % or 1139 hours per year at LB and BB respectively, and 1.04% or 91 hours at PB. At TP1, the seaward edge, there is a notable return to similar tidal conditions across both sites, with an inundation period of 31.8% or 2785 hours per year at BB and 32.1% or 2803 LB. At PB, TP1 between the tidal flat and marsh seaward edge was inundated 8.1% of the year or 708 hours per year, which is more similar to the inundation period of zone 2 in the other two sites, and notably dominated by the same genus as the other two sites.

3.3.4 Low mid and high marsh divisions

Across the whole marsh, natural groupings are evident in the inundation data, and these support the division of a low, mid, and high marsh, where the low marsh (zones

1 and 2), experience a range of 32 % to 9% inundation period, the mid marsh (zone 3), 8 % to around 1%, and the high marsh (zone 4 to 5) 1% to 0.01%. Despite the different tidal range and patterns between the micro tidal sites and the meso tidal site, (see Figure 3.6) the elevation ranges of zones are similar. Zone 1 and 2 which make up the low marsh display the most variation in the inundation period experienced at each transition point.

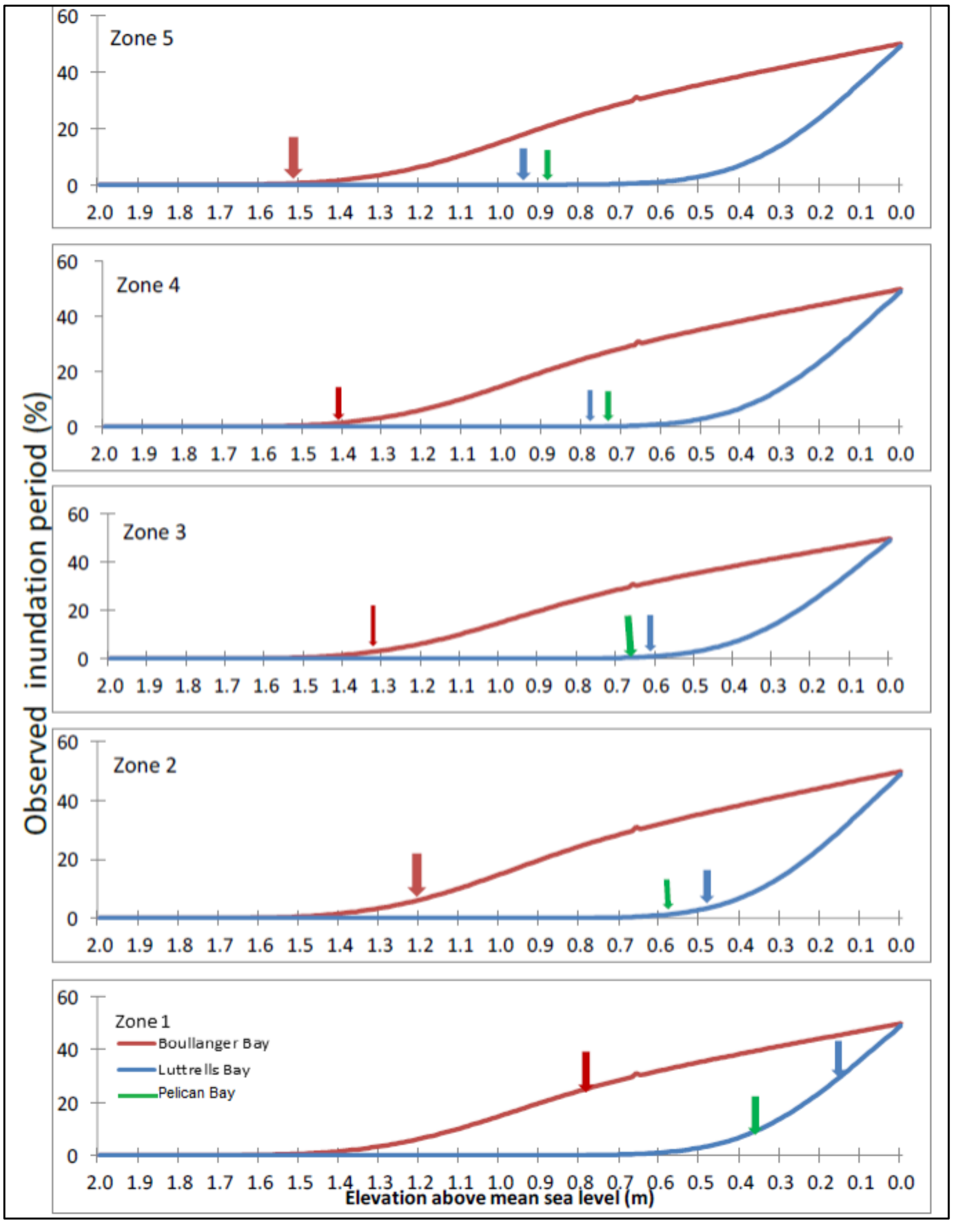


Figure 3-6 Tidal inundation curve for Boullanger Bay, Luttrells Bay and Pelican Bay. Elevation relative to mean sea level is indicated on the x axis, and % inundation period on the y axis with transition points to each zone marked by respective coloured arrows. The difference in inundation period across the three sites is greatest at the seaward edge zone 1, and declines landward. The difference in inundation period for zone 5 transition point at all sites was minimal with less than 0.01% at Luttrells Bay and Pelican Bay and less than 0.02% at Boullanger Bay. Inundation period for both micro tidal sites remained at less than 1% and was 1.3% at Boullanger Bay at the zone 4 transition point.

3.3.5 Multiple analysis of variance

There was a significant difference in soil salinity and pH at the transition points (across the three marshes (see Figure 3-4), ($F=5.796$, df Pillai 40. 7, df = 8, $p = < 0.001$). Transition points 1, 2 and 3 were not significantly different to each other for either pH or salinity but were to zone 4 which decreased for both ($p=<0.05$). There was a significant difference in salinity and pH between TP 4 and 5 ($p = <0.05$), where again, both were higher (see Figure 3-7).

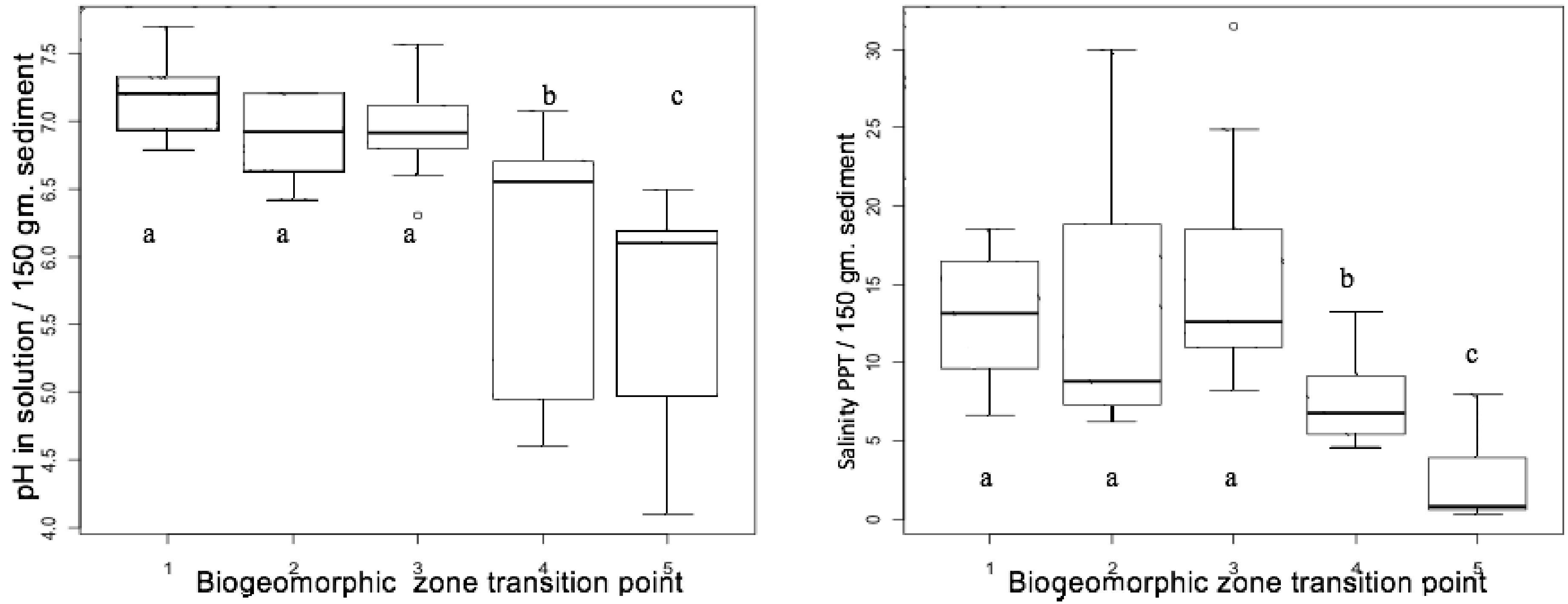


Figure 3-7 Multiple anova results for mean pH and salinity at transition points. There was no significant difference in the mean soil pH and salinity across sites for the low marsh zones 1 and 2, and the mid marsh zone 3. However, they were significantly different to zone 4 which was also significantly different to zone 5. Means with the same letters indicate zones that are not significantly different (p value cutoff = 0.05).

3.3.6 Soil Texture

In Luttrells Bay transects 1 and 2 on the leeward side of the marsh, contained more sand than transects 3 and 4 on the windward side of the marsh. Tidal flat sediment texture was different at each site with silty clay loam and silt at Luttrells Bay, loamy sand at Pelican Bay and fine sand at Boullanger Bay. At TP1 all samples were of sandy clay loam except at transects 3 and 4 Luttrells Bay.(Table 3-4). Relative sand content was less at TP2 compared to TP1 at all sites. Relative clay content was greater at TP3 than TP2 at Luttrells Bay and Boullanger Bay but at Pelican Bay both were clay loam. Texture at TP4 was most similar between sites where there was a return to sandy sediment. The trend continued landward where at TP5 at Luttrells Bay and Pelican Bay sand was dominant, but TP5 at Boullanger Bay contains a higher loam content.

Table 3-3 Soil texture at each site and transition zone transition points (n=50), including the tidal flat showing a high degree of similarity within sites, compared to the high variation of the tidal flat. PU refers to physionogmic units.

site	Tidal flat	TP1	TP2	TP3	TP4	TP5
LB	(PU2)Silty clay loam (PU1)Silt	Sandy clay loam Silty clay loam	Silty loam	Silty clay loam	Sandy clay loam	Loamy sand
PB	Loamy sand	Sandy clay loam	Clay loam	Clay loam	Sandy clay loam	Loamy sand
BB	Sand	Sandy clay loam	Silty loam	Silty clay loam	Sandy clay loam	Sandy loam

3.3.7 Predicting marsh divisions

Soil salinity was able to predict the presence of low marsh at all sites (see Table 3-4). At Boullanger Bay and Pelican Bay it was the only significant predictor. For every 1% increase in soil salinity there was an increase of 1.06 in the likelihood of low marsh presence ($G^2 = 26.57$, df, 1, 37, p = < 0.001) at Boullanger Bay and a 1.21 increase at Pelican Bay. Neither soil moisture, pH or inundation period significantly contributed to the probability of low marsh being present at these sites or at Luttrells Bay. At the latter, both salinity and moisture content were independently statistically significant predictors, and for every 1% increase in both units, there was a 1.1 increase in the probability of low marsh being present (G^2 = 6.89, df, 38, $p<0.01$, G^2 = 8.21, df 38, p = <0.01 respectively).

When data sets for each site were combined to test the generality of these results across regions and tidal ranges, the predictive ability of salinity increased, and inundation period significantly contributed to the predictive ability of the model. For every 1% increase in salinity there was a 1.14 increase in the odds of low marsh being present (G^2 = 1.14, df 100, p = < 0.001). Across sites, for every 1% increase in inundation period, there was a 1.03 increase in probability of low marsh being present (G^2 = 3.89, df 99, p = <0.05). Sediment moisture and pH did not significantly add to the probability of low marsh presence across sites.

The strongest statistically significant predictor of mid marsh presence at Luttrells Bay was sediment moisture content, where for every 1% increase there was an increase of 1.22 in the odds of mid marsh being present (G^2 = 8.21, df 38, p = < 0.01). Inundation period was weakly significant when associated with moisture content, and contributed to the predictive ability of the data set where for every 1% increase in inundation period there was an increase 0.81 in the odds of mid marsh presence (G^2 = 0.5, df 33, p = < 0.5).

At Pelican Bay inundation period was the strongest predictor of mid marsh presence where every 1% increment there was an increase of 0.3 in the odds (G^2 = 18.50, df 32, p = < 0.001), followed by an increase of 1.18 with every 1% increase in moisture content (G^2 = 5.79, df 31, p = < 0.05). The logistic regression for the mid marsh at Boullanger Bay showed inundation to be the strongest significant predictor (G^2 =14.6, df 32, p = < 0.001), in association with moisture content (G^2 = 4.56, df 34, p = <0.05). In association with inundation period and moisture content, salinity at Boullanger Bay was significant at the p = < 0.5 level, but when present in the goodness of fit test, made the test statistically significant, indicating a difference between the observed and predicted likelihoods, and was thus removed from the model.

For the mid marsh the combined data set showed sediment moisture content to be the strongest predictor (for every 1% increase, the odds of mid marsh presence increased by 1.5), in association with inundation period (for every 1% increase the odds increased by 0.33). Across regions, neither salinity nor pH contributed to the probability of the mid marsh being present.

The high marsh could not be modelled for either Luttrells Bay or Pelican Bay individually where both failed the goodness of fit test. However, at Boullanger moisture content was the strongest predictor, followed by inundation period. Neither salinity nor pH contributed to the probability of high marsh being present at a site or regional level. When combined each individual independent variable contributed significantly to the model.

Table 3-4 Multiple logistic regression results for: 1) the likelihood ratio test statistic (G^2), which if significant makes the model a better fit than without it, 2) the goodness of fit (GoF) le Cessie-van Houwelingen normal test statistic (z) for the unweighted sums of squared errors and its probability (non-significance indicates there is no evident lack of fit of the model), and 3) the odds ratio (OR) factor by which the predictor variable increases or decreases the log odds of the response variable (in this case marsh division presence). Blue cells are the independently significant variable and uncoloured are significant when in association with the former.

Site		**Likelihood ratio test**			**GoF**		**OR**
Division	**Predictor**	**G^2**	**df**	**p**	**z**	**p**	
Boullanger Bay							
Low	Salinity	26.6	37	<0.001	2.02	0.44	1.06
Mid	Inundation	14.6	32	<0.001	0.9	0.36	1.3
	Moisture	4.6	34	<0.05			0.7
High	Moisture	21.3	35	<0.001	1.9	0.06	1.5
	Inundation	3.0	34	<0.01			0.33
Luttrells Bay							
Low	Moisture	8.2	38	<0.01	2.8	0.06	1.14
	Salinity	6.9	38	<0.01			1.03
Mid	Moisture	10.9	34	<0.001	0.4	0.67	1.22
	Inundation	0.5	33	< 0.5			0.8
High	Algorithm failed to converge						
Pelican Bay							
Low	Salinity	14.7	24	<0.001	1.2	0.23	0.1
Mid	Inundation	18.5	32	<0.001	2.3	0.29	0.03
	Moisture	5.8	31	<0.05			1.18
High	Algorithm failed to converge						
Combined/Regional							
Low	Salinity	1.14	100	<0.001	1.9	0.6	1.14
	Inundation	3.89	99	<0.05			1.03
Mid	Inundation	20.04	105	<0.001	2.6	0.17	0.86
	moisture	6.92	104	<0.01			1.06
High	Inundation	45.4	106	<0.001	0.8	0.89	0.02
	Salinity	23.9	102	<0.01			0.52
	Moisture	8.21	104	<0.01			0.98
	pH	2.90	104	<0.1			1.05

3.4 Discussion

3.4.1 Predicting marsh zones

Inundation period is the most consistent predictor of marsh divisions at a general level but locally, edaphic factors are more influential. The relative influence of inundation period and edaphic factors can be interpreted from a zones lower elevation across sites, and explained by regional climate and, or geomorphic setting. The most obvious effect of geomorphic setting is in the low marsh where edaphic factors are influenced by sediment characteristics, as evidenced from the tidal flat textures. In the soft sediment environment of Luttrells Bay, moisture content was the strongest predictor. In the sandier, lower estuarine and open coast setting of Pelican Bay and Boullanger Bay, soil salinity was stronger. The result demonstrates an important relationship between predictors and their geomorphic setting that the MLR has detected, because at the soft sediment site, the seaward edge occurred lower in the tidal frame than at the other 2 sites. In combination, the longer inundation period, and finer sediment texture would create high moisture content, and lower salinity (Maricle and Lee, 2007).

The sediment texture and lower elevation of TP1 at Luttrells Bay is also consistent with the relationship described in section 3.1.1, between bottom shear stress as a function of elevation or water depth and consequent wave energy, controlling the distribution of intertidal zones (Fagherazzi et al., 2006). Shear stress thresholds of the intermediate tidal flat increase with finer sediment texture so the transition between the intermediate zone and salt marsh edge can occur lower in the tidal frame (Fagherazzi et al., 2006), see Section 3.1.1.

Vegetation composition of the seaward edge provides further evidence for wetter edaphic conditions at Luttrells Bay where *S. repens* was dominant. In comparison, the seaward edge at the two higher and sandier sites where dominated by *S. quinqueflora*. The absence of pioneering *Sarcocornia* at waterlogged sites is consistent with Northern Hemisphere findings that *Sarcocornia perrenis* is outcompeted in heavily waterlogged conditions (Davy, 2006). A threshold for inundation period is evident for the two *Sarcocornia* spp. At TP2 at Luttrells Bay *S. blackiana* became dominant, at the same elevation and inundation period as *S. quinqueflora* at TP1 at Pelican Bay, and similar inundation period to TP1 at Boullanger Bay. This is also consistent with previous studies demonstrating differential pioneer elevations within marshes or regions for different species (Fagherazzi et al., 2006).

For the mid marsh, biogeomorphic zonation across sites was most strongly predicted by inundation period. The mid marsh consisting solely of biogeomorphic zone 3 *J. kraussii* sedge land, showed very similar elevation range between LB and PB (see Figure 3-2), but its lower extent experienced a five times greater inundation period at Boullanger Bay. At the two sandier sites inundation period was the strongest predictor in association with moisture content.

At Boullanger Bay, TP3 was inundated more frequently than at the other two sites but salinity was lower in the low marsh, as expected with higher annual and growing season rainfall, and lower temperatures, (see section 3-1-3). Interactions between these covariates are likely to be enabling a lower elevation limit at this site. It is consistent with evidence that *J. kraussii* seedling emergence is more successful in lower salinities, even when exposed to daily tides (Engels et al. 2011). Studies from the northern hemisphere show that the lower limit of *J. kraussii* in a saltmarsh is mediated by abiotic factors, and that increased flooding duration decreases salinity tolerance. (Rand, 1976; Naidoo and Kift, 2006). In Tasmania, *J. kraussii* is a common estuarine reed and occurs from freshwater to brackish environments (Kirkpatrick and Glasby, 1981).

The weak significance of salinity at Boullanger Bay, and failed goodness of fit in the predictive model, may be a consequence of the direct coastal exposure resulting in greater micro-topographic relief. Such relief and meteorologically-forced inundation would create the heterogeneous edaphic conditions that were reflected in higher species richness, and a mosaic vegetation distribution (see Table 3-2). Therefore, there was likely to be considerable noise in the sample.

Notwithstanding, the primary aim of the analysis was to test predictability, and so the result is accepted for that purpose. Despite the variation in the lowest elevation of mid marsh across sites, when the data set was combined to test for generality, inundation period was strongly able to predict mid marsh presence in association with moisture content.

Inundation period at TP4 differs between sites, from 113 hours at Boullanger Bay, to 10 hours at Luttrells Bay and 1.2 hours at Pelican Bay. These differences reflect variable meteorological high energy event periodicity in each region, and geomorphic setting. The high marsh consisting of biogeomorphic zones 4 and 5 could not be predicted by any variable at Luttrells Bay or Pelican Bay individually. However, in contrast to the southern drier sites, soil moisture content was strongly able to predict high marsh presence at Boullanger Bay, in association with inundation period.

Species response to local climate and soil types of the landward edge at each site is the most likely explanation for the predictive variability in local data sets. At Boullanger Bay, zone 5 was dominated by *M. ericifolia*, a species that tolerates freshwater to saline environments, and prefers moist soil conditions. This species commonly occurs in mono-specific stands known as coastal paperbark swamp, in regions of high rainfall and poor drainage (Kirkpatrick, 1991). However, at Luttrells and Pelican Bay, the dominant grasses and sedges making up zone 4 were also dominant in zone 5. The most dominant species *Poa labillardieri* occurs from the coastline to the alpine zone in Tasmania (Kirkpatrick, 1991). It could be argued that this zone marks the beginning of the terrestrial transition zone, where flooding stress is eased and salt-intolerant terrestrial species out-compete the halophytes. However, the common presence of this saltmarsh vegetation type in different climatic and edaphic conditions shows that it is a marsh constituent, representing the perceivable limit of tidal influence along the elevation gradient. The different high marsh vegetation types in zone 5 across the region shows that site level predictability of the upper marsh requires subjective definition of its upper limit.

The regional data set was able to predict the presence of high marsh where all predictor variables added significantly to the model. Inundation period was the strongest, supporting the widely held generalisation that tidal inundation is the most important control on salt marsh extent. Salinity, followed by moisture content, were also significant and only in the high marsh across sites did pH contribute to the model, although its significance was low. However, the result is consistent with results for the MANOVA. Pillai's Trace test showed that zones 4 and 5 were significantly different to each other and to zones 1, 2 and 3. Percent moisture content was also able to predict high marsh presence across regions, and sediment texture analysis shows a higher sand content than the lower elevation zones.

3.5 Conclusion

The results of the present study are consistent with those from the well-studied marshes of the northern hemisphere, in that soil salinity and moisture content exert the strongest control on vegetation zonation, (Maricle and Lee, 2007). The interaction between inundation period with sediment texture, and its effect on soil moisture content, is empirically supported and shown to limit tidal flat colonisation and progradation of the low marsh. Soil salinity, in association with inundation period, limits the lower extent of the mid marsh. Inundation period in association with moisture content, controls the lower limits of the upper marsh, and its upper limits are controlled by infrequent, meteorologically forced tidal inundation.

In micro tidal settings, Tasmanian salt marsh seaward edge elevation range is variable, from just above MSL in fine sediment settings when *S. repens* is present, to above mean high water neap in coarse sediment settings when *S. quinqueflora* is present. In the higher meso tidal range, marsh seaward edges occurred lower than mean high water neap. The mid marsh is variably inundated across sites at its lower elevation range, but highly similar across sites at the higher end of the elevation gradient. The high marsh exists variably above mean high water spring at all sites, and thus largely relies on meteorologically forced tidal inundation to maintain its position relative to the tidal frame.

An unexpected finding in this study is that all three marshes occupy similar elevation ranges of around 0.8 m (including the terrestrial transition zone 5), despite being located in places with different tidal ranges. This is not consistent with the prevailing view that the larger the tidal range, the greater the vertical extent of the marsh (McKee and Patrick, 1988; Adam, 1990; Saintilan, 2009; Kirwin and Guntenspergen, 2010), where seaward edges generally start above mean high water, and extend to the highest high water mark. However, as a consequence of the variation in the tidal patterns between the micro tidal and meso tidal sites the relationship between biological zonation and tidal inundation is more easily compared by the graphic representation presented by Ranwell et al. (1964) of critical inundation periods for biological zones indicated on the sigmoidal tidal curve (Figure 3-4). It highlights that studies seeking to contribute to unravelling the complexity of marsh form and process beyond widely held generalities, should express their results in terms of inundation period, rather than as constrained within a certain tidal plane or elevations. When results are expressed as inundation period, sites within different climatic, tidal and geomorphic settings can be compared meaningfully, to properly evaluate diagnostic features from regional data sets for sea level studies.

The transition points between biogeomorphic zones that can be identified in the sedimentary record in saltmarshes were shown to be variable between sites, and dependent on climatic and geomorphic setting. Too great a variation in the range of environmental conditions and controlling factors in biogeomorphic zones may create difficulties in interpreting stratigraphy and fossil species composition. In this instance, suitable variability for a modern analogue may be better achieved by multiple transect sampling or more replicates across key transition points, such as the seaward edge to low marsh, and, mid to high marsh. The suitable regional range may be indicated by vegetation species composition in settings with polyspecific communities such as those in Tasmania, and other south eastern Australia sites.

Chapter 4 A mid Holocene sea level record for the Little Swanport Estuary, south eastern Tasmania

4.1 Introduction

Holocene sea level history for the east coast of Australia has been studied concurrently with the geomorphic evolution of key regional landforms (Thom and Short, 2006). For northern Australia, this includes the evolution of macro tidal estuaries with extensive mangrove system deposits providing evidence for the Last Marine Transgression culminating at 7400 ±200 cal. yr BP, with evidence of higher sea level at 6400 cal. yr BP of approximately + 2.5 m that fell to present level after 1000 cal. yr BP (Mulrennan and Woodroffe, 1988; Woodroffe et al., 1993; Lewis et al., 2013). For Queensland, sea level history was determined from coral micro atolls revealing a mid-Holocene highstand ranging from +1.3 to +1.5 m above present mean sea level (PMSL) between 6770 and 5750 cal.yr BP, falling to present levels around 2000 yr BP (Chappell et al., 1982; Chappell, 1983; Leonard et al., 2016). For New South Wales, a morphostratigraphic approach to understanding estuarine evolution has provided geochronological constraint of modern sea level attainment between 7900 and 7700 cal.yr BP, reaching a highstand of +1 to 1.5 m, between 7700 and 7400 cal.yr BP, which lasted until 2000 cal.yr BP, after which seas fell to their present level (Jones et al., 1979; Chapman et al., 1982; Thom and Roy, 1985; Sloss et al., 2007; Lewis et al., 2013). However, the Holocene sea level history and geomorphological evolution of the Tasmanian coastline is still poorly understood (Clarke et al., 2011).

Recent rates of sea level rise were determined by Gehrels et al. (2012) using the transfer function approach. An increased rate of sea level rise as a consequence of anthropogenically induced global warming was observed. Rates rose from 0.7 ± 0.6 mm / yr, to 4.2 ± 0.1 mm / yr between 1900 and 1950. However, there is little understanding of how the increased rate compares with longer term sea level change and its impact on coastal geomorphological evolution.

The evidence for Holocene sea level change in Tasmania is sporadic and conflicting (Murray-Wallace and Goede, 1995; Haworth et al., 2002; Clark et al., 2011: Gehrels et al., 2012). Earlier research focussed on identifying the existence of a mid-Holocene sea level highstand or sea level oscillations, identifying a sea level of 0.5 ± 0.25 m above present mean sea level at around 2330 cal. years BP (Haworth et al., 2002). Later, Clarke et al., (2011), also suggested a fall in sea level as the most likely cause of reduced marine conditions in the stratigraphic record in coastal waterbodies between ~ 2700 and 1300 cal. yr BP in the south east. For both, the timing is consistent with sea level fall from a highstand on the Australian mainland between 2000 – 2500 cal. yr BP (Sloss et al., 2007; Switzer et al., 2009; Lewis et al., 2013). However, Murray-Wallace and Goede (1991 & 1995) found no evidence for a higher than present sea level from sea level index points (SLIPs), collected to constrain neo-tectonic activity. The uncertainty can be explained by interpretation of fixed ecological indicators exposed to highly variable environmental conditions over time. These fixed environmental indicators are now accepted as marine or terrestrial limiting data points (Sloss et al., 2007; Lewis et al., 2013).

Implicit in sea level studies is the broader concept of coastal evolution (Reading, 1986). Studies concerned with the geomorphic evolution of Tasmania’s Holocene coastal landforms have drawn conclusions on sea level history. Beach ridge systems on the north and east coasts of the island display down stepping patterns (Davies, 1959; Davies, 1961; Jennings, 1961; Chick, 1970; Bowman, 1986 & 1987; Clarke et al, 2011). When direct evidence of sea level change is lacking, such landforms are acknowledged as evidence of sea level regression (Schofield 1960, Harvey, 2006; Ribolini et al., 2011).

Early evidence suggested that the beach ridges had formed some time since sea level reached present levels ± 1 m, around 6000 yr BP (Colhoun, 1983), indicating higher sea levels than present (Murray-Wallace and Goede, 1991). Constraint of the chronological development for Rheban Spit beach barrier system are provided by Bowman (1986 & 1987), who shows four sets of disconformable beach ridges that systematically reduce in elevation by around one meter, from the oldest landward ridge dated at 5500 years BP, followed by ridges at 4200 and 3100 years BP, with the youngest ridge developing sometime in the last century. However, rather than invoking higher sea levels, sedimentological evidence of seaward fining led to the conclusion that the down stepping pattern was a result of gradual reduction in sediment availability over time (Bowman, 1986). This conclusion is consistent with the prevailing view that sediment supply from the continental shelf along south eastern Australia was at first plentiful, but declined over the later stages of the Last Marine Transgression (Nichol et al., 1994). Another major coastal process likely to contribute to seaward decline in beach ridge height over time is that with sea level transgression the near shore will fill with sediment, thus waves fetch and height declines, thereby reducing the elevation at which sediment can be tidally deposited (Doyle et al., 1995). These morphodynamic processes, offer an alternative explanation for such landforms in Tasmania, otherwise interpreted as deposited by higher than present sea levels.

Further south than Rheban Spit, at Bruny Island and South Arm, Clark et al. (2011), as part of a paleo – tsunami study, include biostratigraphic evidence of a transition from freshwater to marine, returning to freshwater conditions, in waterbodies associated with transgressive beach systems. By correlating broad scale coastal change between three coastal water bodies in the area, the simplest conclusion put forward was a late Holocene sea level fall over four phases of landform development, related to Holocene relative sea level change.

None of the existing research prior to recent sea level reconstruction (Gehrels et al., 2012) provide precise SLIPs. Dated samples are either fixed ecological indicators from variable environments over time, or obtained from gross sediment stratigraphy. Such uncalibrated descriptions of the fossil features where error cannot be adequately quantified, or is large, remain useful as qualitative directional indicators (Ellison, 1989). Engelhart et al. (2011a) classify samples found in gross sediment stratigraphy as ‘basal’, in comparison with ‘base of basal’ index points that are located on a surface, unaffected by sediment compaction and, therefore, more precise than a basal index point. Due to the wide error ranges assigned to basal SLIPs, Engelhart et al. (2011a), recommend removing non-base of basal SLIPs from sea level history databases, particularly if their purpose is to populate glacio-isostatic adjustment (GIA) models. However, they acknowledge there are few base of basal SLIPs in the sea level data base for the north Atlantic seaboard, and this is also the case for Australia. Table 4-1 provides a summary of the existing evidence for Holocene sea levels for Tasmania and classifies SLIPs directional or precise, terrestrial or marine limiting, basal or base of basal. It includes indirect evidence of past sea levels from coastal landform evolution studies, and shows that most of these are located in high energy, coarse sediment environments.

Table 4-1Compilation of sea level data points for Tasmania. Data points are classified as directional or precise (after Ellison, 1999), terrestrial or marine limiting and basal or base of basal (after Engelhart et al., 2011a). The compilation shows there are few data points offering precision greater than the potential variation in sea level over time required to determine sea level change over the mid to late Holocene.

Author	Location	Facies and geomorphic setting	Elevation (m)	Age (years BP)	Conclusion
Murray-Wallace & Goede, 1991	Five Mile Beach	Clean quartz sand sequence *Phacasoma coerulea*	Up to 1.5 m above HWM, buried by 1.2 m aeolian sand dune.	Holocene indicated by geomorphic setting	**Directional**
	Ralphs Bay The Spit	Marine sand unit underlying a 0.7 m Holocene Midden. *Katelysia scalarina*	Above present sea level	6510 ± 80	**Directional** Facies contained disarticulated and articulated shells. Insufficient elevation data
	Shell Pits Point (Duck Bay)	Tightly compacted mollusc in grainstone underlying a beach ridge mantled by an immature, grey soil, (0.6 m depth). *Fulvia tenuicosta*	Not indicated	7060 ± 100	**Directional** Considering tidal range, present elevation would be around HWM.
	Shell Pits Point	*Glycymeris (Tucetilla) striatularis*	Not indicated	6020 ± 140	**Directional**
	Shelly Beach, Ralphs Bay	+0.3 m thick shelly beach deposit at the base of low cliff in Ralphs Bay, overlain by 0.3 m midden and again by 0.3 m Aeolian sand. *Katelysia scalarina*	Not indicated	2770 ± 110	**Directional** Did not require higher than present sea level for deposition.

Goede et al., 1993	Launceston, Tamar Estuary	Outer edge of estuarine flood plain in shelly sand sediment underlying urbanised area proximal to estuary tidal limit (not known if it is landfill or estuary fill).	Surface of site c. ≤ 1 m above high tide level and sample found at c. 3.6 m depth.	2600 ± 400	**Directional** Insufficient elevation data
Haworth et al., 2002	Three Rivers Creek, King Island	Shingle storm beach	+1.7 m	790 ± 60	**Directional**
	British Admiral Beach, King Island	Cobble shingle storm beach, comminuted mixed species	Above present MSL	4560 ± 120	**Directional**
	Cameron Inlet, Flinders Island	Shell rich muddy sand, back barrier lagoon *Katelysia sp.*	Burial depth 0.3 – 0.6 m, 0.5 m AHD	2060 ± 100	**Marine limiting** Probably low intertidal
	Sandy Lagoon, Flinders Island	Shell rich muddy sand, back barrier lagoon *Katelysia sp.*, *Batilaria diemensis*	Burial depth 0.1 m, 0 m AHD	3600 ± 110	**Marine limiting** Probably low intertidal
	Fotheringate Bay, Flinders Island	Shelly sand, embayment fill *K. scalarina*, *Mactra pura, B. diemensis*	Burial depth 0.1 m, 0 m AHD	5580 ± 80	**Marine limiting** Low intertidal, may be transported
	Trousers Point Beach, Flinders Island	Sand, back barrier lagoon *Katelysia* sp.	Burial depth 0.2 m, 0 m AHD	3580 ± 90	**Directional** Sandy environments shift too readily to establish deposition
	Grassy Harbour, King Island	Tubeworm*, Glycymeris caespitosa* on granite bedrock outside of breakwater reef facing landward.	0.5 ± 0.25 m + PMSL.	2330	**Directional**
Chick, 1970	Ulverstone, Brigadoon shoreline	Fossil shoreline, low beach ridge system above PMSL	+1 m MHHW	Holocene indicated by geomorphologic setting	**Directional** Possible deposition by higher wave energy than present or higher sea level than present
Bowman 1987	Rheban Spit	Four sequential beach ridges with samples obtained from their corresponding swales, decreasing in elevation seaward.	+2.5 m decreasing to 1.5 m above mean sea level	5500, 4200, 3100 and modern.	**Directional**

Clark et al., 2011	Primrose Sands	Charcoal in freshwater peat	-0.95 m MTL	7424-7256	**Directional** No evidence of open marine conditions
	South Neck Beach	Peat	-1.04 m MTL	9888-9541	**Terrestrial limiting**
		Fresh water peat	-0.55 m MTL	1307-1242 1201-1185	**Marine limiting** Intertidal
		Bembicium nanum, Zeacumantus diemensis, Salinator spp. in tidal inlet sediment	-0.85 m MTL	2776-2675	**Marine limiting** Low intertidal
		Anapella cycladae, Ostrea anagassi at or below low tide in low to medium energy environment, possibly transported	-1.18 m MTL	4218-3970	**Terrestrial limiting**
		Charcoal sample in terrestrial weathered dune sand.	-2.21 m MTL	8404-8293, 8260-8250	**Marine limiting** Intertidal
		A. cycladea in upward fining sediment with brackish marine fossil features.	-1.15 m MTL	5022-4822	**Marine limiting** Intertidal
		Well preserved bivalve shell fragments in brackish tidal inlet.	-1.70 m MTL	6155-6314	
Gehrels et al., 2012	Little Swanport Estuary	*Spisula trigonella* found in vertical position within upper drowned river valley salt marsh, estuarine infill, and assigned an indicative meaning based on low tidal flat heights.	- 0.34 ± 40 m	5677 ± 38	**Base of basal**
		Mactra pura in vertical position 119 cm depth in core	-0.40 ± 0.50 m	4414 ± 37	**Basal**
		Filamentous algae, elevation and foraminifera transfer function.	- 0.27 ± 0.1 m	650 ± 115	**Precise**

Precise sea level indicators have been identified in depositional salt marsh environments of micro tidal estuaries. A saltmarsh foraminifera elevation transfer function (Callard et al., 2011) reconstructed sea level history for the last 650 years BP (Gehrels et al., 2012), predicting MSL with an accuracy of 0.6 m ± 0.1 m. Reconnaissance of Tasmania's saltmarshes revealed no greater marsh depths (unpublished data), and by comparison to older northern hemisphere salt marshes, are unlikely to provide sufficient sedimentary record to reconstruct Holocene sea levels.

Microtidal estuaries are key coastal landforms in Tasmania, particularly along its highly indented south eastern coastline (Kirkpatrick and Glasby, 1981). They have been classified for management purposes using physico-chemical attributes (Edgar et al., 1999), but knowledge of their structure and the nature of their infilling is lacking in comparison to other south eastern Australian regions, where a geochronological framework for estuarine evolution with sea level change has been proposed as a regional model (Sloss et al., 2004; 2006a, b). Such a framework enables integration of regional scale investigations of coastal history to address questions of coastal functioning at extensive spatial and temporal scales (Devoy et al., 1996). In Australia an integrated approach is necessary, because of the evident spatial variability in relative sea levels during the Holocene from the observational record (Chappell et al., 1982; Chappell, 1983; Hopley, 1983; Harvey, 2006; Lessa and Masselink, 2006; Sloss et al., 2007; Switzer et al., 2009), and predicted variation according to the far field glacio–hydro-isostatic adjustment model (Nakada and Lambeck 1989; Lambeck and Nakada, 1990).

The glacio-hydro-isostatic sea-level model (Lambeck and Nakada, 1990), provides RSL predictions from the onset of ice melt at 18,000 years BP. Ice volumes for the Arctic and Antarctic were sufficient for sea levels to be lower than present by about 130 m. After18,000 years BP sea level rose gradually with ice melt, and RSL model predictions show dependence of the water load term on ocean geometry and distance to the shoreline. According to Earth rheological processes, this combined effect causes variable tilting of the continent ocean interface (Whitehouse, 2009; Engelhart et al., 2011; Gehrels and Long, 2008). Predictions of relative sea level over this period show little spatial variation from Cape York to Hobart, with the southern-most location most closely following the eustatic sea level of the transgression (Lambeck and Nakada, 1990). Modelling illustrated that during the period of sea level rise the water volume is the predominant process affecting RSL, but upon ice melt cessation, time dependent processes of isostatic adjustment to the water load become dominant.

The variation in isostatic response by latitude is largely a consequence of variation in the width of the continental shelf, although coastal geometry can affect RSL, particularly in Queensland and South Australia (Lambeck and Nakada, 1990). The combination of factors affect continental levering processes, and has a smaller effect of ocean siphoning with Antarctic glacio-isostatic uplift upon ice unloading. This process causes ocean water to drain from lower to higher latitudes when seabed depression follows the forcing of mantle material from beneath the ocean crust to the lithospheric crust. Collectively these processes cause a relatively small amplitude mid-Holocene highstand, varying in time and space along the eastern Australian margin (Nakada and Lambeck, 1990). Figure 4.1 compiles the sea level envelopes for each region from Lewis et al. (2013), showing the diminishing amplitude of the mid Holocene highstand with latitude, except for South Australia. GHIA modelling shows the variation in South Australia to be a consequence of the water load term on variable shelf geometry for that location, (Lambeck and Nakada, 1990). The glacio–hydro-isostatic adjustment model predicts no highstand for Tasmania (Nakada and Lambeck, 1989) but the conflicting observational record highlights the need for precise SLIPs for that location (Nakada and Lambeck, 1989; Lambeck and Nakada, 1990; Lambeck, 2010).

Greater latitudinal variation is predicted for modern sea level attainment, occurring when meltwater addition from the great ice domes ceased, nominally at around 6000 yr BP (Lambeck and Nakada,1990). However, when adjusted for the ice load, the water load, and the rigid Earth terms (that include upper and lower mantle viscosity and lithospheric thickness), there is a north- south gradient in highstand at this time, declining in magnitude from around 1.5 m in north eastern Australia, to negligible in Tasmania (Lambeck, 2010)

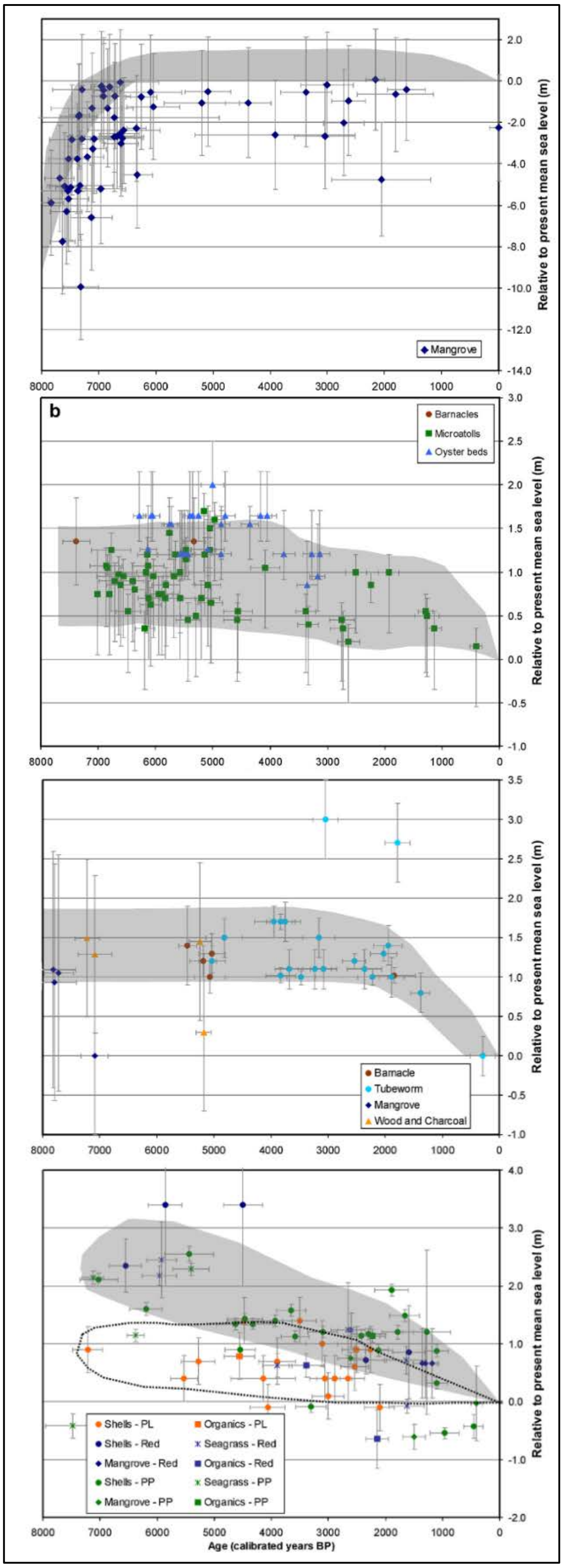


Figure 4-1 Holocene sea level history from a) northern Australia, b) Queensland, c) New South Wales and d) South Australia, showing variation in the timing and magnitude of the mid Holocene highstand along the east coast of Australia. The grey envelope encloses data considered most reliable by Lewis et al. (2013).

Source: Modified from Lewis et al. (2013).

Reliable SLIPs for the more southern locations are required to further develop the GIA model for the Australian continent. Better data will constrain the contribution of meltwater from Antarctica to the global oceans. Geophysical modelling and observations continue to support the proposition of continued slow addition of meltwater from Antarctica (Nakada and Lambeck, 1989; Lambeck and Nakada, 1990; Lambeck and Chappell, 2001; Okuno and Miura, 2013) but the relative contribution to global eustatic sea level is still a subject of debate (Lambeck and Nekada, 1990; Lambeck, 1997; Shennan and Horton, 2002; Peltier, 2002; Milne et al., 2005; Long et al., 2012; Lambeck et al., 2014). Regional geophysical models that account for the contribution of rheological adjustment of the Earth to changing ice/water volumes are important for managers of climate change (Gehrels and Long 2008), who need location specific predictions of sea level change.

The present study aims to; develop a sea level curve for Tasmania with sufficient precision to resolve controversy over the presence or absence of a sea level highstand in Tasmania; add to the sea level index data base for Australia for further iteration of the glacio-hydro-isostatic adjustment model; and, to do so in a manner consistent with the Australian approach of morphostratigraphic investigation of coastal evolution with sea level change and the geochronological framework of estuarine evolution and sea level history in tectonically stable regions recommended by Sloss et al. (2010). The application of the inductive model of Dalrymple et al. (1992) to predict the location of palaeo-environmental zones within the Little Swanport Estuary was a necessary precursor to addressing this main aim.

4.1.1 Study Site

Luttrells Bay within the Little Swanport Estuary is a site with a naturalness index of 2.38 for the estuarine catchment area and 2.74 for its estuarine drainage area (calculated from a 1 to 10 point index of Tasmanian estuaries by Edgar et al. (1999)), indicating a natural condition, but not pristine. Apart from infrequent visits from the landowners the bay has not been disturbed since grazing was restricted in the early 1980s (C. Dyke 2005, pers. comm.). *Spartina alterniflora* populations have been removed and the site continues to be monitored for any reinvasion.

The catchment of the Little Swanport estuary is dominated by Jurassic dolerite which produces a brown nutrient rich clay soil (DPIWE, 2004). The Little Swanport Estuary is elongated and irregularly shaped with a narrow, permanent opening to the ocean (DPIWE, 2004). The river that flows into it is an intermittent watercourse that only flows during rainfall events. Luttrells Bay is the largest bay within upper reaches of the estuary (see Figure 4-2).

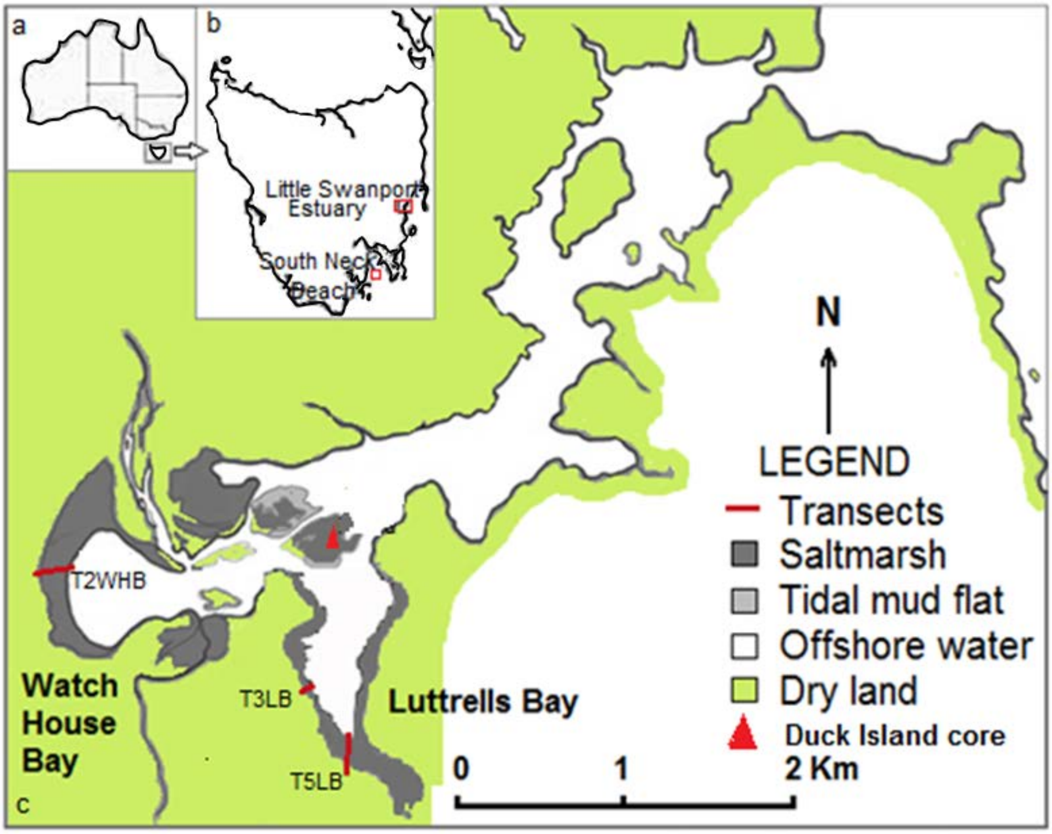


Figure 4-2 (a) location of Tasmania, (b) location of sites mentioned in text, (c) Transect 3 and Transect 5 in Luttrells Bay (T3LB and T5LB), and the location of Watch House Bay and Transect 2 (T2WHB), sampled by Gehrels et al. (2012). Coring locations were all taken along T5LB except for the Duck Island core.

4.2 Methods

To achieve the overall objective of obtaining sea level index points from a single location, in a region where sea level history and its associated stratigraphy is poorly understood, methods were developed in stages and undertaken systematically. Firstly, the estuary structure and its energy regime was examined (4.2.1) and elevation control established (4.2.2). Two coring campaigns were undertaken; the first to describe and characterize the lithostratigraphic and macro biostratigraphic nature of the sedimentary record to ensure its tidal environment of deposition (4.2.4). The characterization was then used to locate an unambiguous transgressive surface for the second, more intensive transect sampling campaign for geochronology (4.2.5). For the latter, samples were only taken for carbon dating if the transgressive surface was unambiguous. Thus coring continued along the transect until a satisfactory range of basal sample elevations were obtained.

4.2.1 Establishing estuary structure and palaeoenvironmental zones for study site selection

Sampling site selection commenced with application of an inductive model (Dalrymple et al., 1992), to establish estuary structure and sedimentary environments using aerial photographic imagery. Reconnaissance by boat was also undertaken for ground-truthing. Subtidal sediment was collected with a hand operated grab-sampler from the three inferred estuarine energy zones (Figure 4-3).

4.2.2 Elevation control and surveying

Two transects were deployed perpendicular to the shoreline within Luttrells Bay. Both were tied to mean sea level (AHD), using survey control points installed within Luttrells Bay from analysis of one differential GPS base station deployed at a location of known height (located 3.87 km away from survey control point 1), and one tied to tidal height (located within 10 m of survey control point 1), hence a baseline length of 3.87 km. Both base stations received satellite data for one hour. Heights relative to Australian height data were resolved from the Os Pos solution. Transects were surveyed and flagged for every 2 cm rise in elevation from the tidal flat, to the marsh upland border after Gehrels (2008). Transect 5 Luttrells Bay (T5LB) was surveyed to characterise the elevation gradient, and to determine the height above mean sea level of the coring locations. The surveyed transects, including those from a previous study (Morrison, 2006), were examined for geomorphological evidence of a higher than present shoreline, including the existence of raised estuarine mud, above the present highest tide mark, that may indicate a higher sea level, consistent with a mid-Holocene sea level highstand.

4.2.3 Coring and Stratigraphy

Fifteen cores sites were identified in two phases. Phase one was conducted to fully characterise sediment facies. Four cores were taken until refusal using a hand operated Hiller Piston Corer with a 1 meter long, 5 cm diameter barrel. The depth of penetration was recorded and checked against the retrieved core length for compaction or disturbance during extrusion from the barrel into clean, PVC pipe, halved lengthways. These cores were then visually logged for colour using Munsell Soil Color Charts (2000), sediment texture and inclusions in the field. Cores were wrapped and stored for transport to an X-ray facility within 24 hours.

One of these cores was TCLB-1 taken at 0.43 m above MSL from the upper end of the lower salt marsh along Transect 5 (T5LB). This core was correlated with C2LB taken from 0.33 m above MSL at the lower end of the low marsh. MFLB-1 was taken from the intertidal mudflat at the seaward end of T5LB at an elevation + 0.11 m relative to MSL. Another core was taken in the upper estuary on Duck Island, to check whether it had evolved as a bay head delta, or whether it was a remnant island, cut off from the shore by recent sea level rise. Ecological indicators such as the dominance of a late stage succession salt marsh species (*Tecticornia arbuscula*) (Prahalad et al., 2011; Prahalad et al., 2015), suggested that it was older than the shoreline fringing marshes within bays. Either landform type would record the timing and magnitude of sea level change. However, an understanding of its evolutionary pathway could inform the direction of sea level changes and enhance the interpretation of the main cores from Luttrells Bay. The elevation of the Duck Island core was 0.62 m above MSL. Each of these cores are described in detail in Section 5.5.3.

4.2.4 Stratigraphy

To characterize the environment of deposition of each facies, lithostratigraphic and biostratigraphic analyses were undertaken. Facies were systematically subsampled for grain size analysis at the termination of, and, prior to the facies contact, and where changes in mollusc abundance were identified from core x-rays. Sediment samples of 4 cm depth were extracted from the core tube, leaving a film of sediment that had contact with the tube. Intact gastropods and articulated and disarticulated, but not broken, bivalves were removed, and their depth in core recorded.

To disperse the fine particles in the remaining sediment each sample was added to 300 ml of sodium meta-hexaphosphate solution for 24 hours, and then placed in an automated shaker for 10 minutes. The samples were left to settle for 48 hours and the fluid was decanted through 25 μm filter paper. The wet sediment, including shell hash, was dried at 60°C for 48 hours, weighed, and separated into size fractions using cascade sieves with apertures of 2000, 1000, 500, 300, 125 and 63 μm. The sieves were shaken for 15 minutes on an automated sieve shaker and the dried sediment on the filter paper, and the separated size fractions were then weighed. Grain size parameters were calculated in Microsoft Exel using the GRADISTAT Version 8.0 grain size analysis program (Blott, 2010). Samples were then examined under a dissection microscope for grain shape after Powers (1953), and the relative proportion of each mineral component, plant fragment and shell hash was determined with the aid of “Charts for Estimating Proportions of Mottles and Coarse Fragments” in Munsell Soil Color Charts (2000), see Table 4.4. Above each sampled location, a second sample was taken for percent organic carbon by loss on ignition, after Heiri et al. (2001). Sediment containing less than 10% organic content was considered inorganic.

To broadly characterise the intertidal environment of deposition of each facies, a third 2 cm sample was taken and wet sieved through 250, 125 and 63 µm sieves for foraminifera. The 125 and 63 μm fractions were scanned under a compound microscope and fossils were picked and mounted on gummed micro-palaeontology slides. Unless otherwise stated, species were identified with the aid of Hayward et al. (1999), and Yassini and Jones (1995). Their preferred habitat ranges are described after the same authors in facies descriptions, see Sectio4.3.4. Descriptions of the observed taxonomic features used to classify species against the literature are in Appendix 1. Molluscs species were classified with the aid of eastern Australian mollusc online databases which also described their habitat preferences within the tidal frame (Groves, 2015: Beechey, 2017). The Zoology Department at the Museum of Tasmania assisted with identification of species whose diagnostic features were damaged by weathering. Table 4-3 lists the species present and the height at which they were first located down core, and dominant species ranges are indicated in stratigraphic diagrams in Section 4.3.4.

4.2.5 Geochronology

A second set of eleven cores were taken using a Hiller barrel corer to retrieve samples for carbon dating, because this corer minimises compaction and core disturbance. The purpose was to retrieve base of basal samples after Engelhart et al. (2011a). In the context of the current study, basal samples are those containing material from the base of the estuarine sediment facies, directly overlying the inferred transgressive surface, observed from the Phase 1 cores. Coring was conducted along T5LB until core refusal and by “feel” for a change in sediment density, which generally occurred within a further 10 to 20 cm, after the unconsolidated overlying sediment was fully breached.

Upon opening the barrel of the Hiller corer, the surface of the sediment was removed to avoid contamination, and examined for macro fossils as close as possible to the contact between the inferred transgressive surface and the overlying Holocene estuarine sediment. Where no fossils were evident, approximately 5 grams of sediment, equating to a 3 mm vertically thick segment was removed from the centre of the core, using a cleaned stainless-steel spatula. The samples were wrapped in alfoil previously sprayed with methylated spirit, placed in plastic zip-lock bags and sent to the Australian National University Carbon Dating Facility for AMS ^{14}C dating after Fallon et al., (2010). A marine reservoir age of 300 ± 30 years was applied after Stuiver et al., 1998b and was the recommended reservoir correction method for the range of mollusc fossils and bulk sediment samples as recommended by the ANU Radiocarbon dating facility (Dr Fallon and Professor Lambeck, pers. comm) The indicative meaning of all samples was developed from fossil features first, supported by sedimentological evidence, allowing direct comparison between different indicators, see Table 4-5.

4.3 Results

4.3.1 Estuary structure

Clear zonation and geomorphological features consistent with a wave-dominated estuary were observed (Figure 4-3). Whilst there is no barrier typical of wave-dominated estuaries because of the bedrock constrained mouth, features consistent with the wave-dominated model include the flood tide delta, with numerous channels, and wash-over deposits from the fore-dunes at the mouth. Other diagnostic features evident in aerial photographic analysis included the muddy depocentre that is well defined in the middle low energy zone. In the upper fluvial-dominated zone there is an extensive, well-developed bay head delta. Further evidence for the classification is provided by the sediment texture analysis showing a decrease in sand content in each energy zone, and an increase in silt and clay in the upper estuary in each energy zone (see Table 4-2).

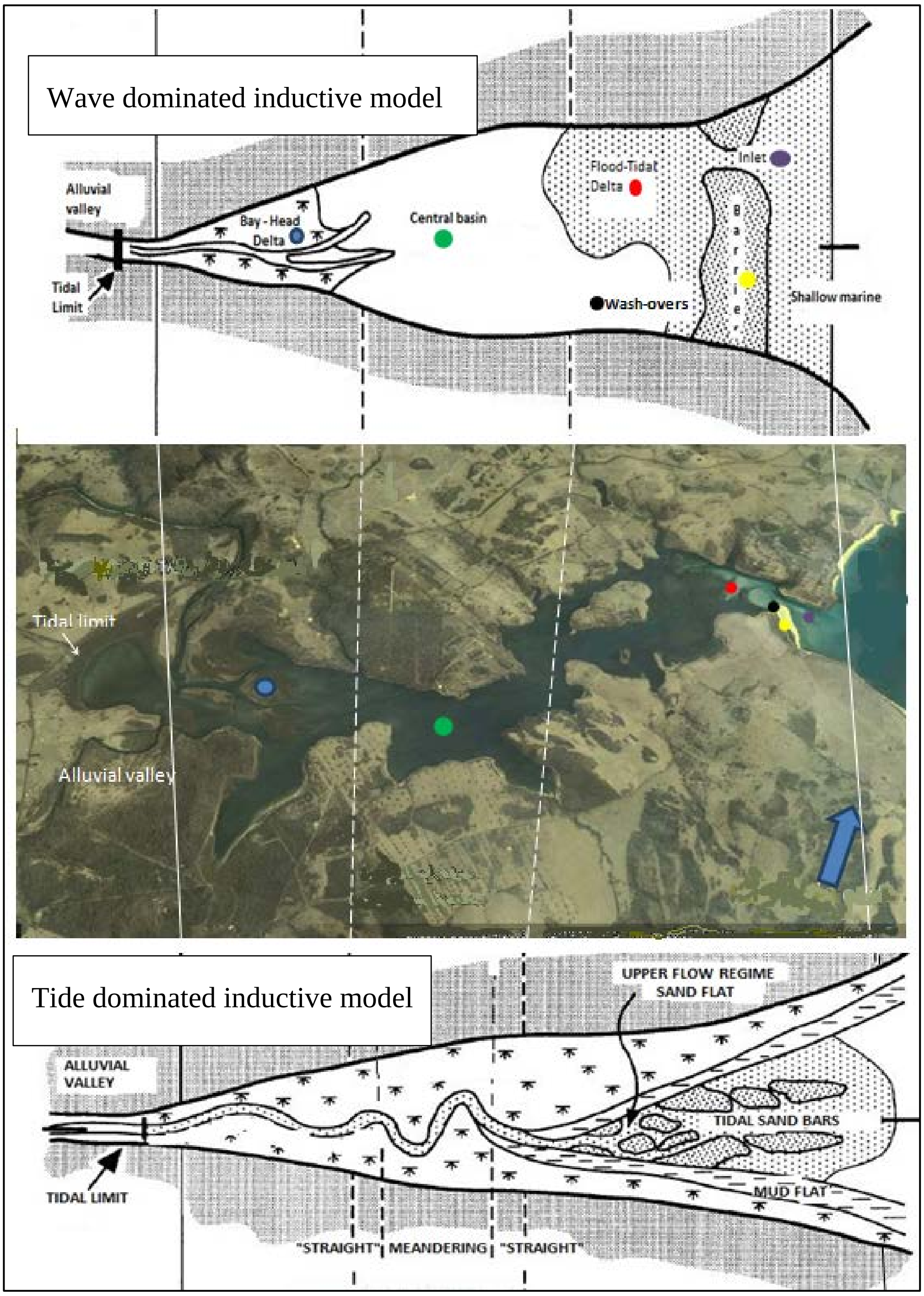


Figure 4-3 Simple comparison of inductive wave and tide dominated models after Dalrymple et al., 1992, and Google Earth image of the Little Swanport Estuary. Considerable similarities are evident between the aerial photo and the wave-dominated model (dots of the same colour indicate comparable locations), but little between the study area and the tide-dominated model.

Table 4-2 Geomorphic features of the inductive and actualistic model of Little Swanport Estuary showing how the estuary conforms to the wave dominated model.

Feature	Inductive model		Little Swanport actualistic model	Sediment / macro-fossils	Result
Dominant energy	Wave-dominated	Tide-dominated	Micro tidal, moderate wave energy (average 1.5 m swell, predominant longshore drift northward).	N/A	Wave/tidal
General whole estuary morphology	Barrier, spit. Straight/ central meandering/ straight morphology.	Incised drowned river valley usually tide dominated. Meandering central region non-existent.	Incised drowned valley, inverted funnel shape by constriction of mouth due to antecedent topography. Prograding beach barrier formation. Straight channel with central meandering.	N/A	Wave/tidal
Tripartite zonation	Well developed, 3 distinct zones.	Less well developed, bay head delta not usually present.	Three distinct zones - well developed bay head delta, muddy central zone and flood tide sand body.	Z 1 coarse Z 2 fine Z 3 mixed	Wave
Outer marine-dominated morphology	Barriers with broad course to med grain sandbars, tidal delta and tidal channels.	Elongate sand bars & broad sand flats passing into low sinuosity single channel.	Broad sandbars dissected by channels, tidal delta and tidal channels. Elongate sand bars not present in constricted mouth but does pass into low sinuosity single channel.	Coarse to medium well sorted sand. Marine shell fragments.	Wave
Low energy central zone	Well-developed muddy with shallow lagoons. Extensive saltmarsh.	Poorly developed, muddy sediments in tidal flats and marshes along margins. May have a sandy channel.	Muddy sediments in tidal flats and marshes in bays along the side of the estuary.	Muddy sand with isolated shoal deposits of estuarine bivalves (>80%).	Wave/Tidal
Inner fluvial-dominated morphology	Bay head delta – tidal influence small.	Inner river-dominated zone has a single, low sinuosity (straight channel).	Well-developed bay head delta. Extensive fringing marshes, one dominant channel, others infilled. Shallow lagoons, tidal channels passing directly into river channel in pro delta region.	Mixed coarse to medium muddy sand, marine/ brackish molluscs.	Wave

4.3.2 Macrofossil analysis

Fourteen mollusc species were identified in the three cores from Luttrells Bay. Mid intertidal range gastropods were dominant in all cores with some low subtidal intertidal species in the lower facies. Intact burrowing shallow subtidal bivalves were less frequent, but all species habitat preferences were limited to sheltered environments such as estuaries (Grove, 2015). TC1LB contained all species represented in other cores (see Table 4-3), that were less diverse. The density of molluscs in the x-ray image of TCLB-1 is typical of the density in all cores, see Figure 4-4 a and b.

Table 4-3 Molluscs in core TCLB-1 listed in order of where they were first located down core, and their preferred habitat and tidal range in Tasmania, according to Grove (2015), and Beechey (2017).

Depth (cm)	**Species**	**Preferred Habitat**
87	*Velacumantis australis*	Sheltered bays and estuaries in sea grasses, **intertidal**
108	*Austrocochlea constricta*	Sheltered rocky sandy shores, in seagrass, **midtidal**
126	*Spisula (Notospisula) trigonella*	In sand and mud in sheltered environs, **subtidal**
	Cryptassiminea tasmanica	Any sediment type but prefers estuarine conditions, **intertidal**
141	*Zeacumantus estuarina*	On mud, in sheltered environs, **intertidal to subtidal**
149	*Nassarius (Niotha) pauperatus*	Amongst seagrass on mud to fine sand, **low intertidal**

154	*Diala suturalis*	Amongst seagrasses in sheltered environments, **intertidal to subtidal**
	Ostrea angassi	Sandy to muddy estuaries, **intertidal**
	Katelysia scalarina	In sand to sandy mud, **intertidal**
	Gaimardia (Neogaemardia) tasmanica	Sandy environs offshore, **subtidal**
160	*Nassarius pauperatus*	On muddy sand, in Tasmania prefers creeks and mudflats, **low subtidal**
	Nassarius nigellus	On or in mud to sand-flats, **low intertidal to subtidal**
	Austrochoclea porchata	Lives on the surface on moderately exposed rocky shores to sheltered sand and seagrass in estuaries, **midtidal.**

4.3.3 Sediment analysis

Other than the change in colour and sediment texture from the salt marsh facies down core, there was very little visually observable change through the rest of the facies in all cores. Grain size analysis showed a predominance of fine sand with varying proportions of silts. The relative proportions of sand and silt produced mean grain sizes ranging from 238 to 77 μm. Minerology was dominated by angular quartz, and was also crystalline in the lower facies, see Table 4.3. The coarse silt sized particles were dominated by dolerite grains that were a blocky shape and the finer silt and clay were amorphous. There was a consistent, but minor presence of amphibole that was prismatic, feldspar that was angular, and biotite that was platy throughout all cores, see Figures 4-5, 4-6 and 4-7.

The minerals were mixed with varying quantities of shell hash in facies below the salt marsh, ranging in size from 0.5 mm to 1 cm (greater than 1 cm was considered broken mollusc). The shell hash also occurred as distinct layers within facies, see Figure 4-5, 4-6 and 4-7. In samples analysed with shell hash retained, mean grain sizes were 460 to 385 μm and were named fine gravelly - fine sand.

Table 4-4 Sediment lithology and approximate % volume, mean grain size, sorting, kurtosis and sediment texture group.

TC1LB						
Depth in core	Sediment composition	Mean grain size μm	Sorting	Skewedness	Kurtosis	Sediment group
10-14	60% plant 5% biotite 35% quartz, angular	102	Poorly sorted	Fine Skewed	Leptokurtic	Fine sandy silt
30-34	50% Quartz, 80% angular and 20% rounded. 30% plant 10% amphibole 10% amorphous -brown	118	Poorly sorted	Fine skewed	Leptokurtic	Fine sandy silt
48-52	50% Quartz, angular 30% Plant 20% amorphous silt	77	Poorly sorted	Fine skewed	Very leptokurtic	Fine sand silt
56-60	30% Quartz, angular 30% Medium shell hash 20% amorphous silt 10% amphibole 10% dolerite	186	Poorly sorted and bimodal			Silty fine sand
70-74	35% shell hash 60% angular quartz 5% biotite	148	Poorly sorted	Fine skewed	Leptokurtic	Fine sandy silt

TC1LB						
120-124	50% quartz, angular and crystalline 40% medium to fine shell hash 5% brown amorphous 5% dolerite < 1% biotitie and feldspar	238	Moderately sorted	Coarse Skewed	Leptokurtic	Fine gravelly fine sand
140-144	30% medium to fine shell hash 50% quartz, angular and crystalline 10% brown amorphous 7% dolerite 3% biotitie and feldspar	159	Poorly sorted	Fine skewed	Leptokurtic	Fine silty fine sand
164-167	20% fine shell hash 60% quartz, angular and crystalline 10% dolerite 5% brown amorphous 5% biotite and feldspar	186	Poorly sorted	Fine skewed	Very leptokurtic	Coarse silty fine sand

TC2LB						
15-19	50% quartz, angular 40% plant 10% biotite, prismatic	98	Very poorly sorted	Very fine skewed	Leptokurtic	Fine sandy silt
30	50% quartz, angular 40% plant 10% amphibole, prismatic	107.3	Very poorly sorted	Very fine skewed	Leptokurtic	Fine silty fine sand
60-64	75% quartz, angular 15% dolerite, blocky 10% amphibole, prismatic	150.9	Poorly sorted	Fine skewed	Very leptokurtic	Silty sand
75a	40% medium shell hash 40% quartz, angular, crystalline 10% dolerite, blocky 10% amphibole, prismatic <1% other (biotite)	386	Poorly sorted and bimodal			Very fine gravelly fine sand
75b shell hash removed	10% fine shell hash 60% quartz, angular, crystalline 20% dolerite, blocky 10% amphibole, prismatic <1% other (feldspar, biotite)	176	Poorly sorted	Fine skewed	Leptokurtic	Silty sand
115	80% quartz, angular, crystalline 15% dolerite, blocky 5% amphibole, prismatic <1 other (feldspar, biotite)	167	Poorly sorted	Fine skewed	Very leptokurtic	Coarse silty fine sand
160 without shell hash >500 *u*m	10% fine shell hash 80% Quartz 8% biotite 2% other	235	Poorly sorted	Fine skewed	Leptokurtic	Fine to medium sand

TC2LB						
160 - 164	40% coarse shell hash. 35% quartz, angular and crystalline 3% other (feldspar, biotite)	385	Poorly sorted	Fine skewed	Leptokurtic	Fine to medium sand
180 - 184 without shell hash	20% fine shell hash 65% quartz angular and crystalline 10% dolerite 5% other (feldspar, biotite)	176	Poorly sorted	Fine skewed	Leptokurtic	Fine sand
180 shell hash &sediment	40% coarse shell hash 35% quartz, angular and crystalline 10% dolerite 3% other (feldspar, biotite)	460	Poorly sorted	Fine skewed	Very leptokurtic	Very fine gravelly fine sand

4

MFLB-1							
10	25% fine shell hash 50% quartz angular 15% dolerite 7% brown amorphous 3% biotite, prismatic	159	Poorly sorted		Very leptokurtic	Silty fine sand
28	60% quartz angular 15% dolerite 10% fine shell hash 10 % brown amorphous <5% biotite and feldspar	176	Poorly sorted	Fine skewed	Very leptokurtic	Silty fine sand
60	40% fine to medium shell hash 40% quartz angular and crystalline 10% dolerite 5% brown amorphous	323	Poorly sorted and bimodal			Fine gravelly fine sand
83	60% quartz angular and chrystalline 20% dolerite 10% fine shell hash	162	Poorly sorted	Fine skewed	Leptokurtic	Silty fine sand
158	60% quartz angular and crystalline 20% dolerite 10% fine shell hash 3% biotite prismatic 2% feldspar angular < 5%brown amorphous Small pebbles	205	Poorly sorted	Very finely skewed	Leptokurtic	Very coarse silty fine sand

4.3.4 Core stratigraphy

TCLB -1

Facies 1 Saltmarsh, fine sandy silt, 0 – 0.46 m

The saltmarsh facies was dark reddish brown fine sandy silt. At 0.10 and 0.30 m depth in core, the mean grain size was 102 μm and 118 µm respectively, and both were fine skewed and leptokurtic. However, microscopic observation showed a significant proportion of root material was retained in each sieve and thus included in the analysis result. Plant material was greatest in the 500 μm fraction, making up approximately 40% of the sediment fraction.

In situ the sediment was bound by the dense and extensive *Sarcocornia blackiana* root mat, with no visible deterioration in the rootlets throughout the facies. The density of the mat decreases below 45 cm depth. Organic carbon content was 29% in the top 0.01 m, 20% at 0.1 m, 21% at 0.3 m, rising again to 29% at 0.4 m and 26% at 0.43 m.

Molluscs were absent from the salt marsh facies. Salt marsh foraminifera were abundant (n=52), and dominated by *Trochaminna inflata* at 10 cm depth, with a relative abundance of 91%. *Jadaminna macrescens* and poorly preserved *Quinqueloculina seminula* were also present with a relative abundance of 3% each. Other species included *Haplophragmoides wilberti* (2%) and *Elphidium excavatum clavatum* (1%). All species present are commonly found in association in salt marshes (Hayward et al., 1999)

Facies 2 Organic, fine sandy silt, 0.46 – 0.58 m

A sharp contact occurred at 0.46 m with a change to black, sandy silt and a reduction in *S. blackiana* rootlets. At 0.5 m the mean grain size was 77 µm, and was poorly sorted, fine skewed and very leptokurtic. As per the declining root content trend, organic carbon content dropped consistently below 46 cm depth with 17%, 15%, 12% and 10% organic carbon at 0.46 m, 0.49 m, 0.52 m and 0.55 m respectively. The occurrence of a shell hash layer at 0.51 – 0.57 m contributed to a sediment mean grain size of 186 µm. This poorly sorted, bimodal layer precluded the calculation of skewedness and kurtosis. *S. blackiana* rootlets were still present and molluscs were absent from facies 2. Foraminifera were dominated by broken *Ammobaculites exigus* (83%), which is a low intertidal to shallow subtidal species, mostly intact *Trochammina inflata* (9% relative abundance) and *H. wilberti* (8%). Rare, and broken fragments of *Milliamina fusca*, a low marsh to low intertidal species (Hayward et al and Yassini and Jones), were evident but were not counted in the population census as they were less than approximately 30% intact and also dissolved on picking.

Facies 3 Inorganic, fine sandy silt, 0.58 – 0.87 m

A gradational contact occurred between 0.55 and 0.60 m through to a very dark grey, fine sandy silt with an increase in shell fragments (> 35%). At 0.7 m, the mean grain size was 135 μm with a poorly sorted, fine skewed sediment matrix. Organic content at 0.58 m decreased to 8.9% from 10% or above in facies 2, although rootlets were still present. Foraminifera were not common (n=23), and consisted of 95% *A. exigus* fragments and 5% *T. inflata.*

Facies 4 Inorganic, shell rich fine sand, 0.87 – 1.29 m (fine gravelly fine sand)

Facies 4 was determined by a sharp contact to a shell-rich layer. Whilst not visible externally, the x-ray revealed a layer of mixed intact and broken molluscs at 0.87 m. *S. blackiana* rootlets were also faintly visible. At 0.9 m the mean grain size was 135 µm, poorly sorted, coarsely skewed and platykurtic. At 1.2 m the shell hash content increased and the sediment was moderately sorted, coarse skewed and leptokurtic with a mean of 238 µm. The decrease in organic carbon content downward through the core, evident in facies 1, 2 and 3, was not present in facies 4. All samples between 0.60 m and 0.85 m contained less than 7.12% organic carbon (SE ± 2.1%). Below 0.85 m to the bottom of the facies at 1.29 m, the fraction of shell hash increased as did the number of macrofossils, making subsampling for LOI unreliable as an indicator of organic content.

The first mollusc down core at 0.87 m was the relatively large gastropod, *Velacumantis australis* which prefers intertidal habitats in sheltered bays and estuaries, often within seagrasses. The most frequently occurring gastropod was *Austrocochlea constricta* which is known to live in the intertidal zone between the low tide mark and the splash zone, from rocky exposed shores to estuaries and marine lakes. *Cryptassiminea tasmanica* was also abundant, and is known to occur on any intertidal sediment type, although it prefers estuarine conditions and is common in salt marshes. All three species were present on the modern surface at Luttrells Bay on the upper intertidal mudflats, close to the salt marsh edge. The least common species was disarticulated *Spisula trigonella* which lives subtidally in sand and mud in sheltered environments.

Foraminifera at 0.90 – 0.92 m depth in core were rare (less than 10 specimens), and were dominated by *Haplophragmoides wilberti* (75% relative abundance), *Trochammina inflata* (15%) and *Milliammina fusca* fragments which dissolved on picking (10%).

Facies 5 Silty fine sand, 1.29 – 1.67 m

A gradational contact occurred between 1.29 and 1.40 m to a dark grey to black silty fine sand. The mean grain size at 1.4 and 1.6 m were 159 and 186 µm respectively and both were poorly sorted, fine skewed and leptokurtic. The darker colour of the sediment is likely explained by a depletion in shell hash content in this facies. Infrequent rip-up clasts were evident in the form of coarser aggregates of sand and the few foraminifera found at the top of the facies were 100% poorly preserved *Ammobaculites exigus*. At the base of the facies (1.65 – 1.67 m) the few foraminifera present (n=19), were a mixture of the low intertidal species *Ammobaculites spp* (68%), and salt marsh species *Trochammina inflata* (21%), *Trochammina salsa* (5%) and *Jadammina macrescens* (5%). Fragments of *M. fusca* were present but disintegrated when picked.

The most abundant intact mollusc in facies 5 was *Katelysia spisula* which was common in all cores. Broken fragments of *K. spisula* made up much of the shell hash throughout all cores. The species lives intertidally in mud and sand (Groves 2015) preferring the lower littoral zone (Wells and Roberts, 1980; Sloss et al., 2006). Other estuarine intertidal species present included *Cryptassiminea tasmanica* and *Nassarius pauperatus. V. australis* was present throughout the core and *Austrocochlea porchata* was the most abundant at the base of the facies and known to live on sheltered rocky shores, estuaries and marine lakes from low tide to high tide.

Facies 6 Dense mottled clay 1.67 – 1.88 m

A sharp contact occurred at 1.88 m depth at core refusal. A sediment sample was retrieved from the outside of the corer blade in sufficient quantity to do basic texture and colour analysis, showing that the sediment was a dense, sticky white to grey clay. Numerous fragments of plant material were present on the surface, and with aa thick mottled olive green layer. Molluscs and foraminifera were absent.

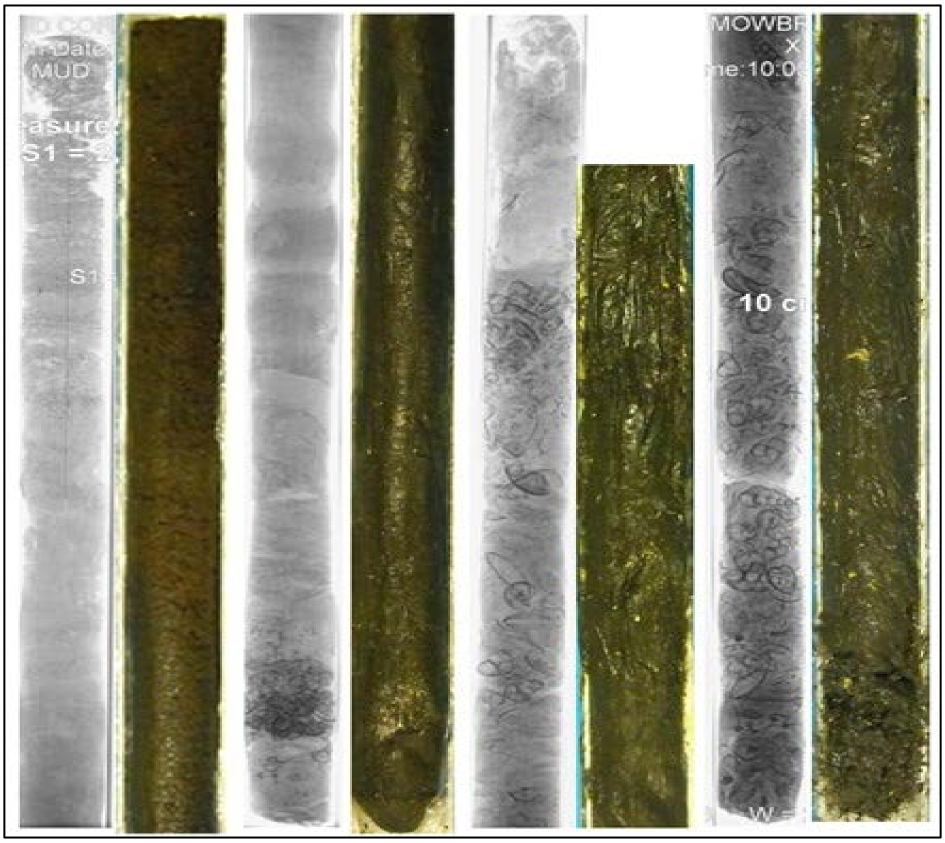


Figure 4-4a Photos of core TCLB-1 and their x-ray images, each length is 0.5 m beginning with 0 at the top left and 2 m at the bottom right. Fourteen cm are missing from the third length.

Figure 4-4b Section from TCLB-1 x-ray at 1.78 to 1.86 m showing intact *V. australis*, *A. porchata* and *K. scalarina*.

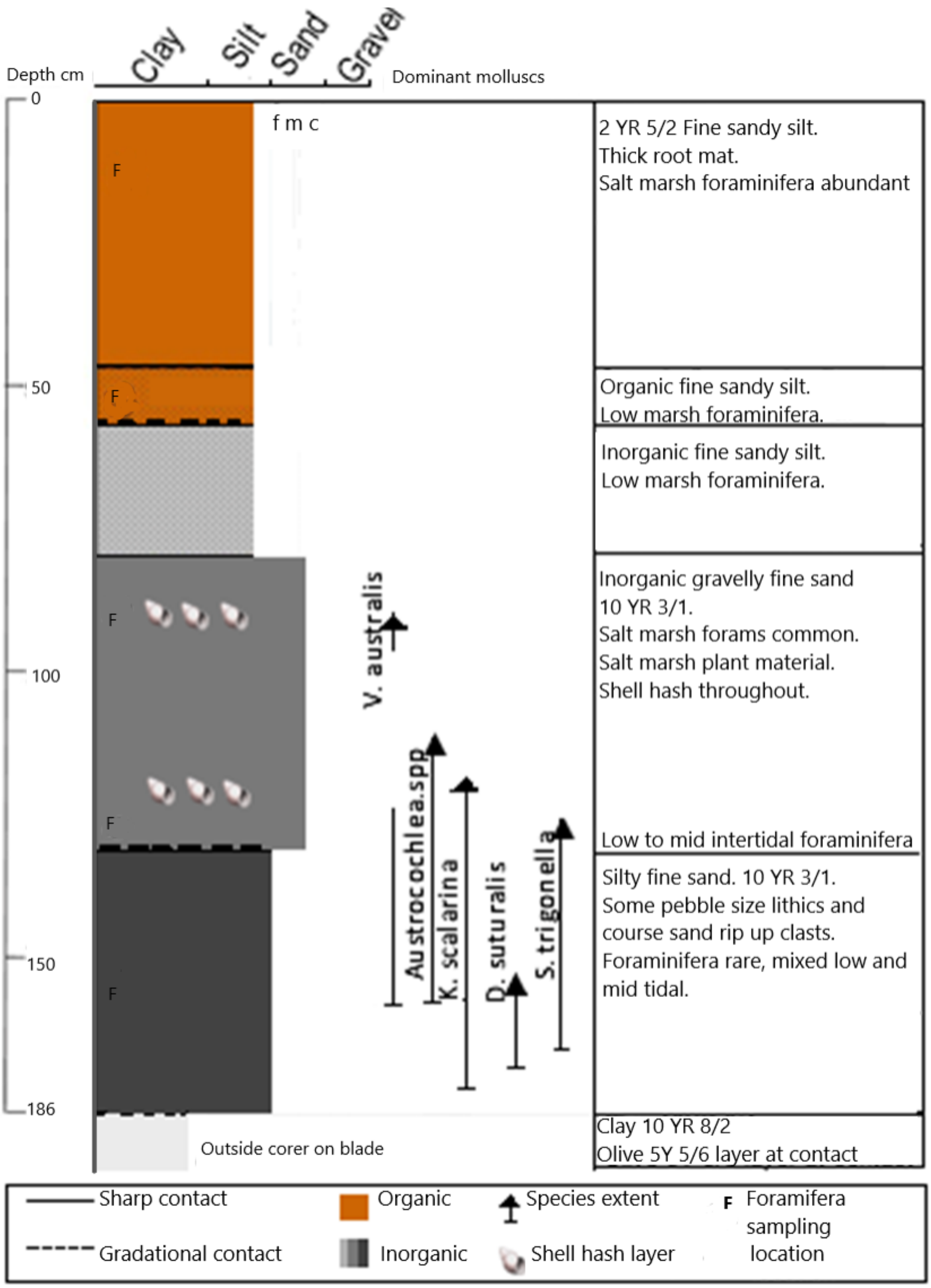


Figure 4-5 TCLB-1 Core stratigraphy including the range of dominant molluscs and tidal range preferences of foraminifera.

TCLB2

Facies 1 Saltmarsh, fine sandy silt, 0 – 0.38 m

The saltmarsh facies was a dark reddish brown silt, containing a small proportion of fine sand. With a mean grain size of 107.3 µm at 0.3 m depth, the sediment was poorly sorted, leptokurtic and finely skewed. At 0.15 m the mean grain size was 98 µm, and was very poorly sorted, very finely skewed, leptokurtic fine sandy silt. Each grain size fraction contained a large proportion of plant material, particularly the 500 µm fraction which incorporated around 50% plant fragments. Organic carbon content was greatest in facies 1 at the top with 29% carbon content at 0.01 m, declining to 20% at 0.1 m, and 21% at 0.3 m. Although molluscs were absent from the salt marsh facies, foraminifera were abundant, dominated by salt marsh species at 0.10 to 0.14 m depth, with a relative abundance of 87% *Trochammina inflata*, 11% *Haplophragmoides wilberti* and 2% *Jadaminna macrescens*.

Facies 2 Organic, silty fine sand, 0.38 – 0.65 m

An 8 cm gradational contact occurred from facies 1 to 2, to a dark grey silty fine sand. Organic carbon content was 21% at 0.40 m, rapidly declining to 13% at 0.52 m and 7% at 0.62 m which was consistent with the observed decline in fine *S. blackiana* roots. The sediment at 0.60 m had a mean grain size of 150.9 µm, was poorly sorted, very finely skewed and leptokurtic. The mineral component of the sediment was composed of 75% quartz with a predominantly angular shape, some prismatic fragments of amphibole and blocky dolerite.

Molluscs were absent from facies 2 and foraminifera recovery was very low. Samples taken at 0.49 - 0.52 m and 0.60 - 0.62 m, yielded a total of 27 foraminifera which were dominated by low intertidal *Ammoculabites exigus* (47% relative abundance), a small round *Ammobaculites sp.* (42%), and broken *Trochammina inflata* (5%).

Facies 3 Inorganic, fine sand(with shell hash very fine gravelly fine sand), 0.65 – 0.80 m

A slight gradational contact occurred between 0.50 and 0.58 m to a very dark grey, silty fine sand with coarse shell hash comprising approximately 20 - 30 % of the sediment. With the shell hash retained, the sediment had a mean grain size of 386 µm, which was poorly sorted and bimodal, thus precluding the calculation of kurtosis and skewedness. *K. scalerina* fragments were the only discernible species observed in the shell hash. When the coarse to medium grain shell hash was removed from the sample, the mean grain size was reduced to 176 µm and was poorly sorted, leptokurtic and finely skewed.

Organic content at 0.68 m dropped to 6.9% with fine *S. blackiana* rootlets still observable. Foraminifera were rare, consisting of 100% broken fragments of *Ammobaculites exigus* . Three intact *Austrochoclea porchata* molluscs were present between 0.68 and 0.70 m and *S. blackiana* rootlets were just visible.

Facies 4 Inorganic, fine sandy silt, 0.80 – 1.14 m

Facies 4 was determined by a sharp contact to very dark grey, coarse silty fine sand which was devoid of shell hash and broken or intact molluscs. The sediment had a mean grain size of 167 µm, was poorly sorted, very leptokurtic, and finely skewed. Thirty-two foraminifera were located at 1.0 to 1.02 m with a relatively even distribution of salt marsh and low intertidal species, *A. exigus* (36%), *T. inflata* (39%) and *H. wilberti* (25%). *Sarcocornia sp*. plant fragments were visible in proximity to the salt marsh foraminifera. Carbon content was variable with 8.3% at 0.9 m and 4.2% at 1.1 m.

Facies 5 Inorganic, shell rich silty fine sand with shell hash (very fine gravelly fine sand), 1.14 – 1.82 m

A slightly visible gradational contact occurred from 1.14 m to a dark grey to black coarse silty fine sand. Grain size was finest at the base of the facies at 1.15 m with a mean grain size of 167 µm which was poorly sorted, finely skewed and very leptokurtic. At 1.60 m the sediment was comprised of approximately 40% shell hash and 35% angular and crystalline shaped quartz grains. Mean grain size was 385 µm and was poorly sorted, leptokurtic and finely skewed. When coarse to medium sized shell hash was removed, the mean grain size decreased to 235 µm and was poorly sorted, leptokurtic and finely skewed. At 1.80 m the sediment with shell hash retained was a very fine gravelly, fine sand, with a mean grain size of 462 µm, was poorly sorted, very leptokurtic and finely skewed. The sediment was comprised of approximately 40% coarse shell hash, and 35% quartz grains that were predominantly angular with crystalline components. After removing shell hash with a grain size greater than 500 µm, the mean grain size dropped to 176 µm and was poorly sorted, leptokurtic, finely skewed, fine sand.

Intact molluscs were abundant (n= 58) between 1.14 and 1.78 m depth in core and consisted of *A. porchata* (72.4%), *K. scalerina* (10.3%), *V. australis* (8.6%), and *D. suturalis* (3.4%). Between 1.3 and 1.5 m, there were 3 intact *S. trigonella*, the only subtidal species, which accounted for 5.2% of the assemblage. Broken molluscs with fragments > 500 µm were frequent and made up 40% of the sediment. Occasional rip-up clasts were evident in the form of coarser aggregates of sand, and there were a few 0.075 m clast. Foraminifera were infrequent at 1.20 to 1.22m with only two species present; the well preserved intertidal *A. beccari* (57%), and the poorly preserved *Ammobaculites* species (43%).

Facies 6 Dense mottled clay, 1.82 – 1.90 m

A sharp contact occurred at 1.74 m consisting of a continuous dark olive layer of around 3 mm depth, to a dense, sticky yellowish grey clay with green mottling. Molluscs and foraminifera were absent.

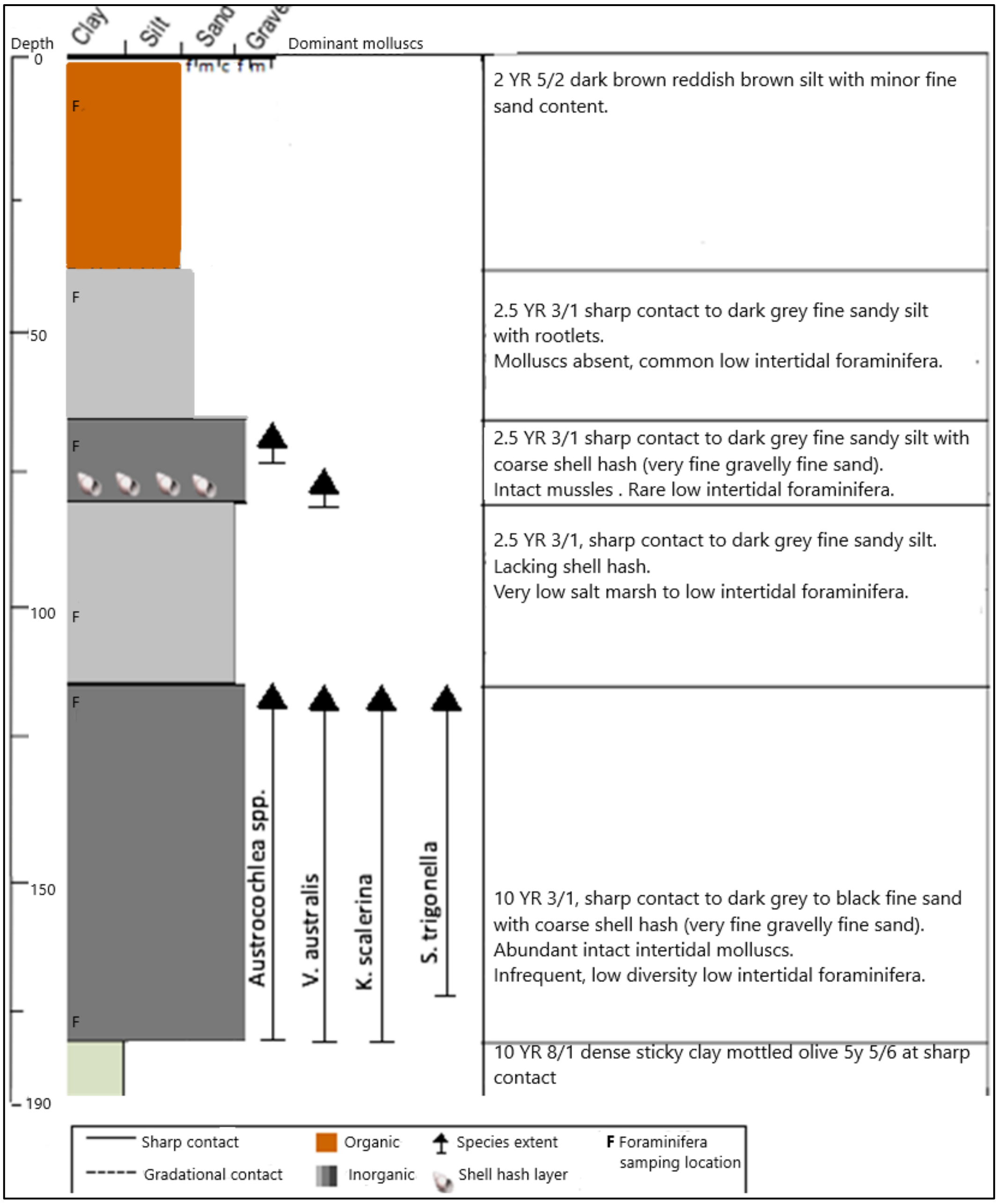


Figure 4-6 Stratigraphic features of C2LB including mollusc species ranges and preferred tidal range of foraminifera.

MFLB -1

A 165 cm core was retrieved from the tidal mudflat (MFLB-1), at the seaward end of transect 5 Luttrells Bay (T5LB) (Figure 4-9). Four facies were evident in the field and laboratory analysis showed the presence of a shell hash layer that was classed as a fifth facies, see Figure 4-8.

Facies 1 Organic, fine sandy silt, 0 – 0.13 m

Facies 1 consisted of black to dark grey silty fine sand, with a mean grain size of 159 μm that was poorly sorted, fine skewed and very leptokurtic. The facies had an organic content of 13 ± 4% (n= 4). Abundant modern foraminifera were present (n=32), indicating low intertidal assemblages dominated by *Elphidium exclavatum clavatum* (80%), *Ammonia aoteana* (14%) and broken *T. inflata (6%).* Molluscs were absent at the top of the core.

Facies 2 Inorganic, fine sandy silt, 0.13 – 0.37 m

A slight gradational contact occurred between facies 1 and 2, to a fine sandy silt. There was no visible change in colour or texture although organic carbon content was less than 10% at 0.20 m depth in core. A distinct mollusc layer was evident in x-ray images at 0.15 to 0.22 m and was comprised of intact, unweathered *A. porchata* with some disarticulated *K. scalerina* and other unidentifiable mollusc fragments. Foraminifera were rare at 0.22 m, (n=14) and consisted of broken *Ammobaculites spp*. and two *Ammonia spp.*

Facies 3 Fine gravelly, fine sand, 0.37 – 0.63 m

A sharp contact occurred to facies three comprised of a fine gravelly fine sand, made up from fine quartz and other mineral components, and shell hash sized between 500 to 1000 μm. At 0.60 m the sediment with shell hash included, produced a mean grain size of 323 μm, was poorly sorted and bimodal, thus precluded further grain size distribution analysis. A distinct mollusc layer was present at 0.60 to 0.63 m, with 5 individual intact *A. constricta* and one intact *V. australis* which would have occupied the same area in core as the five *A. constricta*. There were numerous disarticulated *K. scalerina*. Foraminifera were rare (n=7), consisting of 100% broken individuals of *Ammobaculites spp.*

Facies 4 Silty fine sand, 0.63 – 1.54 m

A gradational contact occurred from facies three to four, to a silty fine sand with less shell hash that was also finer than in facies 3. Intact gastropods and largely disarticulated bivalves were abundant and dominated by *Austrocochlea spp.* that extended throughout the facies at 70.6% relative abundance, *D. suturalis 11.8%*, articulated *S. trigonella* at 8.8% and *K. scalerina at 5.9%*. One intact *Salinator fragilis*, was present (2.9%), whose modern distribution is limited to intertidal mudflat habitats with sea grasses (Hedge and Kriwoken, 2000). At 1.45 m foraminifera were infrequent (n=9), consisting of broken *Ammobaculites sp.* (88%), and, weathered unidentifiable *Ammonia sp.* at 22% relative abundance.

Facies 5 Very coarse silty fine sand, 1.54 – 1.65 m

A gradational contact occurred from 1.54 to 1.65 m to a black very coarse silty fine sand with some small pebble sized clasts and aggregates of quartz sand not mixed with silt that presented as rip up clasts. At 1.58 m the sediment had a mean grain size of 205 µm, was poorly sorted, very finely skewed and leptokurtic. At 1.60 m foraminifera were present (n=14), comprised of largely broken *Ammobaculites spp*. at 78.6% relative abundance, and *Ammonia sp*. at 21.4%.

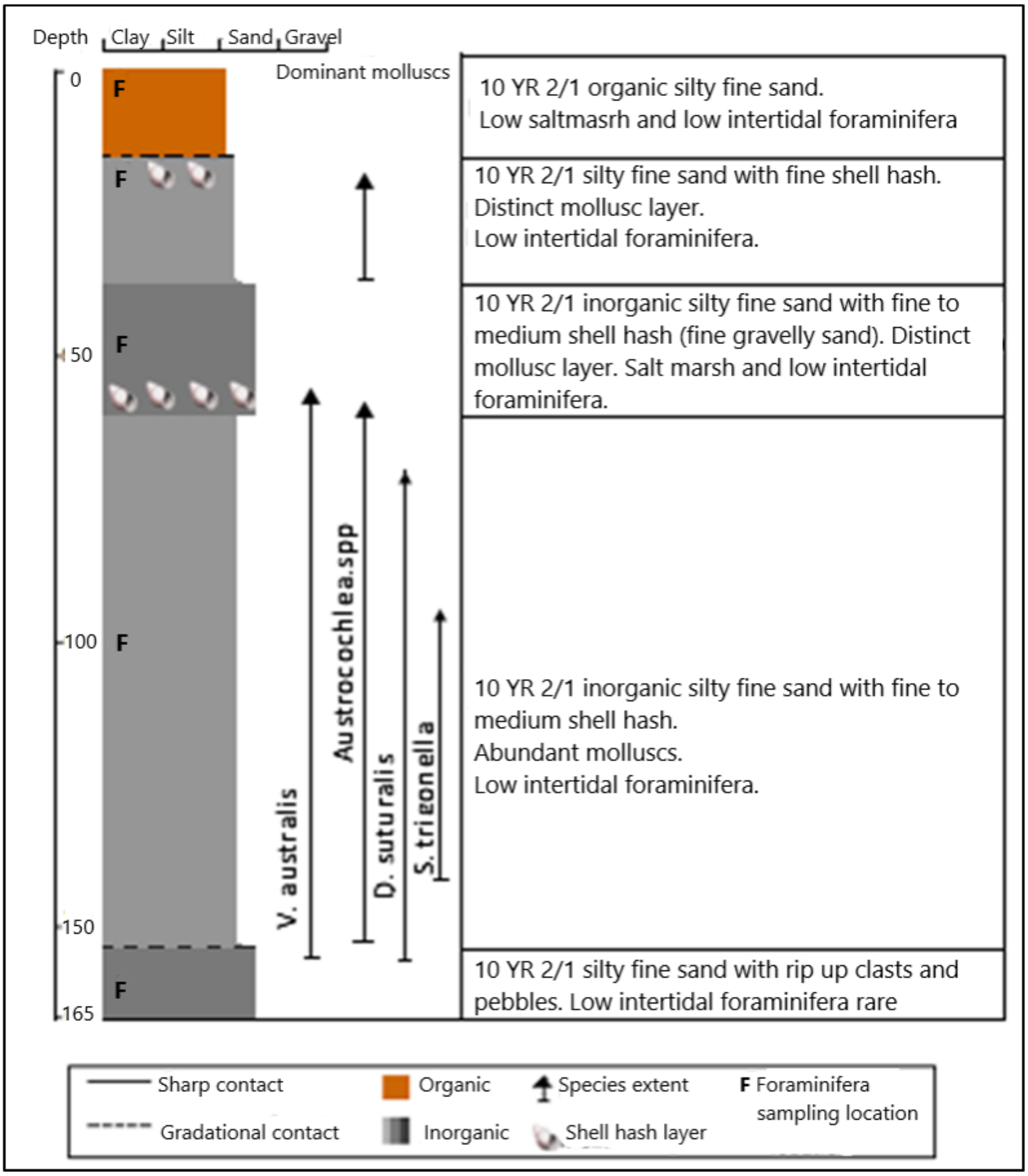


Figure 4.7 Stratigraphy diagram MFLB-1. The colour change in the diagram reflects the variation in the amount of shell hash present in the facies, as opposed to the overall colour of the facies which remained the same throughout the core.

Duck Island

A 1.4 m core to refusal was retrieved from Duck Island. Soft intertidal mud and rising tide precluded the transportation of cores or samples back to the laboratory. Four facies were evident on visual observation in the field, based on slight colour change and the presence or absence of rootlets. The sedimentary sequence from Duck Island differed from the sequence in Luttrells Bay in that underlying the saltmarsh facies, the facies were almost featureless, comprised of silty medium to fine sands, with occasional biological inclusions such as wood and charcoal fragments. Molluscs were absent from the Duck Island core.

Facies 1 Saltmarsh, 0 – 0.38 m

Facies 1 extended to 0.38 m depth. It was reddish brown silt with a small amount of fine sand. An extensive root matt of *Sarcocornia blackiana* bound the sediment.

Facies 2 0.38 – 0.89 m

Facies two was a medium grey colour consisting of fine to medium sand interspersed with rootlets. Biological inclusions included plant material, wood and charcoal.

Facies 3 Sand-rich mud, 0.89m – 1.30 m

Facies 3 was defined by a lack of rootlets and was a medium grey, medium to fine sand with shell fragments present.

Facies 4 Sand-rich mud with molluscs, 1.30 – 1.40 m

Core refusal occurred at 1.40 m in highly compacted sand. Grain size was medium to fine sand with some coarse lithic particles. The sediment was a darker grey than the preceding facies and had a low proportion of shell hash with some broken molluscs present.

4.3.4 Sea Level Index Points

Nine new sea level index points were added in the present study from samples retrieved from the contact between the estuarine sediment and the inferred transgressive surface. Figures 4.8 shows the facies contact that was clearly observed in all cores, and 4.9 shows the positional nature of macrofossils found between the facies. The indicative meaning was assigned as per the species ecological range if it were intact or a large identifiable fragment (Table 4.5), which gave improved chronological results with depth (Table 4.5, 5th column), otherwise the assumption of MSL was applied.

Figure 4-8 Core T5LB9b base showing the section subsampled at the 2cm gradational contact between the dense clay of the inferred transgressive surface (right) and the overlying estuarine sediment facies (left). The two pen marks on the core barrel are 2 cm apart.

Figure 4-9 *Diala suturalis* fragment sample embedded in the contact between the estuarine fine sandy mud and the inferred transgressive surface which was subsampled for carbon dating.

4.3.4.1 Macrofossil indicative meanings

Core 4 on transect 5 (C4LB5), was 1.07 m long, a fossil plant fragment was removed from the sediment directly above the 2 cm contact to the inferred transgressive surface (ITS) at 0.74 – 0.76 m. Core 5 (T5LBC5) located at 0.48 m above MSL ±2 cm, was 1.36 m long. The sharp contact between the ITS and the overlying sandy estuarine mud occurred at 1.28 – 1.30 m depth in core and contained an intact *Velacumantis australis*. Its intertidal range is MSL to mean low tide and thus the indicative meaning constrained to – 0.3 ± 0.3 m relative to mean sea level.

Core 6 (T5LBC6) refusal occurred at 1.83 m with a surface elevation of + 0.44 m relative to MSL. A large piece of bivalve, likely to be *Katelysia scalerina*, was sampled at the sharp contact between the ITS and overlying estuarine sandy mud facies at 1.75 m depth in core. This species lives in mud within sheltered intertidal to shallow subtidal environments (Walton et al., 2002; Grove, 2015) and was assigned an indicative meaning – 0.6 ± 0.4 m. Core 7 (T5LBC7) was located at + 0.2 m relative to MSL and was 3.46 m deep. The inferred transgressive surface was not reached and the lower facies consisted of a mixed but distinct mosaic of yellow brown clay and medium grey muddy estuarine sand. At 2.53 m to 2.45 m a distinct shell layer of mostly shell hash with some disarticulated and broken bivalves occurred. A large fragment of *Diala suturalis* was sampled at 0.87 m depth in core at the contact between dark grey muddy sand and the overlying lighter grey sandy mud. This species lives on mud in sheltered intertidal and subtidal environments amongst seaweed and seagrass (Grove 2015), and was assigned an indicative meaning of - 0.3 ± 0.3 m.

Core 8 was located at + 0.27 m relative to MSL and refusal occurred at 3.78 m. The basal facies occurred at 3.44 m depth in core and was determined by the presence of mixed estuarine intertidal sandy mud and sediment of the same colour and texture as the ITS. There was no clear contact evident however, so it was not sampled for dating.

Core 9a (T5LBC9a), located at + 0.30 m above MSL, was retrieved to a depth of 2.45 m. A 2 cm gradational contact occurred from the ITS at 2.22 m depth in core, from which a very small shell fragment (smaller than shell hash) was retrieved. The fragment was assigned an indicative meaning of MSL ± 0.3 m. Core 9b (T5LBC9b) was located at + 0.39 m relative to MSL and refusal occurred at 2.20 m depth. A large piece *K. scalerina* was retrieved at 1.98 m depth in core, at the top of the 2 cm gradational contact to the ITS that occurred at 2.0 m depth in core. This species lives in soft sediment, intertidal to shallow subtidal, sheltered environments (Grove, 2015). It was assigned an indicative meaning of 0.6 ± 0.4 m.

Core 10 (T5LBC10_3) was located at + 0.38 m relative to MSL and refusal occurred at 2.77 m depth. A sample of broken bivalve was retrieved from the ITS surface, located on the corer blade at 2.67 m. Cores 11_1 (T5LBC11_1) and 11_2 (T5LBC11_2), were located at + 0.50 m and 0.54 m relative to MSL respectively and refusal occurred for both at 1.21 m depth. A sharp contact occurred between the ITS and the overlying estuarine sediment of black fine sandy mud at 1.125 m and 1.16 m respectively. The estuarine facies was devoid of any inclusions such as mollusc fragments or plant material, and so bulk sediment samples were taken directly above the contact and assigned and indicative meaning of MSL ± 0.4 m.

Table 4.5 Samples dated for SLIPs in this study, including core height relative to mean sea level, sample depth in core ± 0.124 m elevation control error, it's indicative meaning, its palaeo sea level position and calibrated ^{14}C results.

Laboratory code and core name	**Core height relative to MSL (H) - core depth** (cm)	**Core height - sample depth relative to MSL ± 12.4 cm elevation control error (D)** (cm)	**Indicative Meaning (I)** (cm)	**Palaeo sea level position relative to present MSL = (H-D-I)** (cm)	**^{14}C age 94.5% confidence limits Calibrated years BP**	**Sample material**
S-ANU36314 T5LBC11_2	50.5 – 116.5 = -66	50.4 – 112.5 = -62.1 ±12.4	0 ± 40	-62.1 ± 52.4	4665 ± 126	Bulk sediment
S-ANU63613 T5LBC11_1	54 – 116= -62	54 – 116 = -62 ±12.4	0 ± 40	-62 ± 52.4	4937 ± 166	Bulk sediment
S-ANU35813 T5LBC9b	39 – 220 = -181	39 – 198 = -159 ±12.4	-60 ± 40	-99 ± 52.4	4905 ± 120	*K. scalarina,* broken piece

S-ANU35810 T5LBC5128	48 - 136 = -88	48 – 128 = -80 ±12.4	-30 ± 30	-50 ± 42.4	5057 ± 137	*V. australis* intact
S-ANU35812 T5LBC7 345	20 – 253 = -233	20 - 87 = -67 ±12.4	30 ± 30	-37 ± 42.4	5324 ± 102.5	*D. suturalis intact*
S-ANU36312 T5LBC10_3	38 – 276 = -238	38 – 276 = -238 ±12.4	-60 ± 20	-178 ± 32.4	5610.5 ± 110.5	*K. scalerina* fragment from the corer blade
S-ANU35811 T5LBC6	44 – 183 = -139	44 - 175 = -131 ±12.4	-30 ± 30	-101 ± 42.4	6996.5 ± 113.5	*D. suturalis*, intact.
T5LBC9a245	30 – 245 = -215	30 – 222 = -192 ±12.4	-30 ± 30	-162 ± 42.4	7363 ± 71	Unidentified shell fragment / hash
SANU35832T5 LBC4	48 – 107 = -52	48 – 74 = -26 ±12.4	0 ± 60	-74 ± 72.4	Modern	Plant

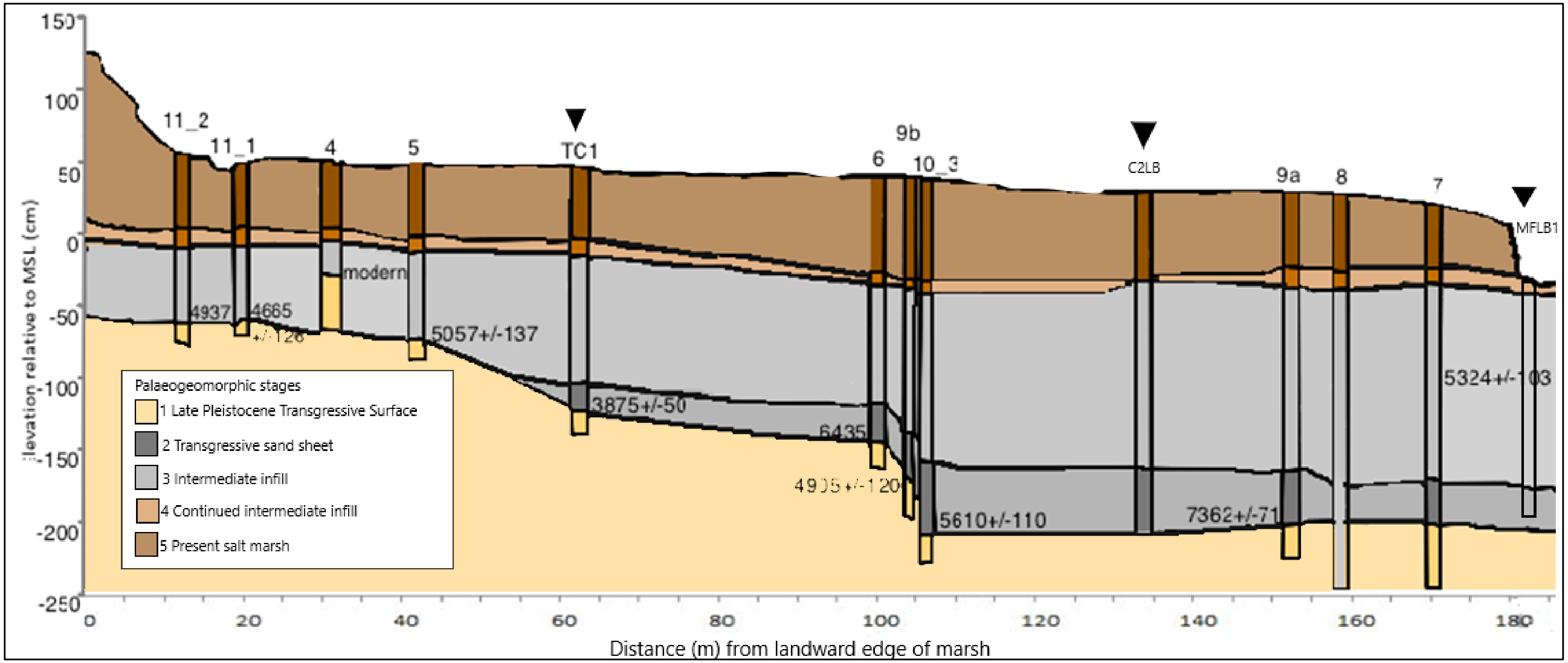


Figure 4-10 Transect 5 Luttrells Bay (T5LB) profile indicating core locations, depth and basic stratigraphy showing the profile of the underlying transgressive surface and depth of the sea level index point samples.

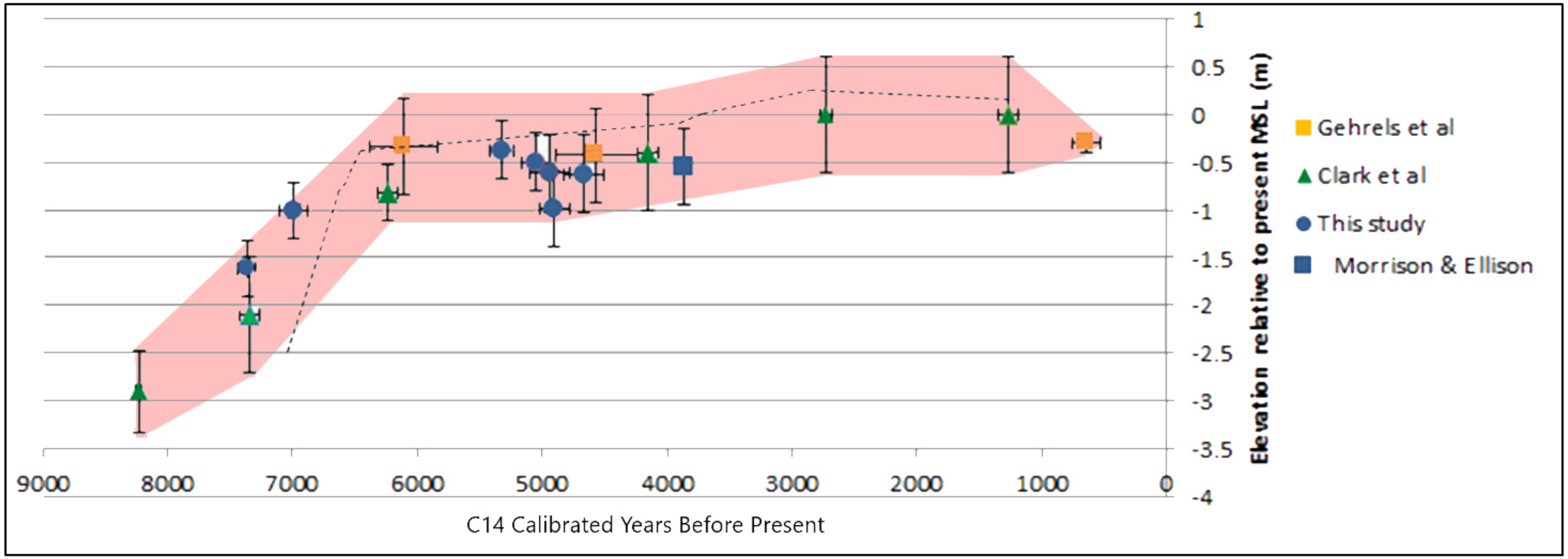


Figure 4-11 A mid to late Holocene sea level curve for south eastern Tasmania showing a rapid rate of relative sea level rise until around 7200 yr BP, slowing at around 6000 yr BP followed by stability, or a very slow rate until 660 yr BP. The black dashed line is the relative mean sea level predictions for Tasmania according to the GHIA model cited from Gehrels et al. (2012). The pink shading shows the vertical error range in this study.

4.4 Discussion

4.4.1 Estuarine structure

The Little Swanport Estuary can be classified as a drowned river estuary that is today a wave-dominated estuary with a bedrock confined mouth that would act similarly to a barrier (Figure 4-3). However, as an antecedent condition, the influence of barrier growth on energy conditions reflected in sediment deposition inside the estuary would not be comparable with the influence described for estuaries in New South Wales (Sloss et al., 2006b). Gehrels et al. (2012), previously described the estuary as tide-dominated, probably because the unusual bedrock confined mouth would limit wave energy entering the estuary, thus allowing tidal energy to predominate. However, based on the evidence provided in the present study, 5 of the 7 key features that distinguish tide- from wave- dominated estuaries confirm it to be the latter (Table 4-2). This conclusion is also consistent with the prevailing view that micro-tidal estuaries are usually dominated by wind and wave action (Perillo, 1995). Indeed, bar built estuaries are frequent in the highly indented but low relief coastline (Kirkpatrick and Glasby, 1981). It is also likely that the wave-dominated processes are enhanced by low rainfall, making river hydrodynamics less influential than tides.

Based on this classification, the appropriate location for coring to obtain a mid to late Holocene record was identified as a quiescent bay with a large depositional environment, indicated by extensive fringing saltmarsh showing clear and consistent vegetation zonation, that was tightly constrained to elevation relative to MSL (see Section 3.4, and displayed a mature evolutionary stage morphology. At this location tidal energy is likely to be dominant over wave energy, and its remoteness from direct fluvial inputs enhances the sea level signal. The consistent leptokurtic grain size distribution dominated by very fine quartz sand, and the mixture of estuarine and marine macrofossils throughout reconnaissance cores in Luttrells Bay confirmed these expectations (Table 4-2).

4.4.2 Holocene Sea level change

The sea level curve (Figure 4.11), includes 10 new base of basal SLIPs, and the sea level data points of Gehrels et al., (2012), and Clark et al., (2011) see Table 4-1. While Clark et al., (2011) is a paleo-tsunami study, their data have been assigned indicative meanings, thus can be incorporated in analyses. The sea level curve provides an adequate number of observations to determine a rate of relative sea level rise of 1.3 mm/year from a position of around 2.7 m below MSL at 8349 yr BP, to around 1.0 m at 6996 cal yr BP. This is consistent with the cessation of melt water addition from the major ice sheets of the northern hemisphere, as the major contributor to RSLR around 7000 cal yr BP (Lambeck and Chappell, 2001). The rate of sea level rise decreases after this time to a rate of 0.49 mm/year, reaching around –0.37 m, relative to PMSL, at 5324 cal yr BP, which is consistent with the cessation of major ice melt from Antarctica at around 6000 years BP (Lambeck and Nakada, 1990). Between this time to around 650 cal yr BP, sea level change stabilises, although a very slow rate of SLR of 0.04 mm per year is the most likely interpretation of the SLIPs and their vertical error margins.

After around 5324 cal yr BP, there appears a slight fall in MSL of around 0.15 m before rising by the same amount by 4905 cal yr BP. However, before invoking sea level oscillation, it is possible that the variation is a consequence of conservative application of the indicative meaning formula, calculated by subtracting the midpoint of a fossil's elevation range, and its upper and lower elevation limits that provide the error ranges. However, the indicative meaning of the bounding SLIPs for this period were calculated from the elevation range of the entire tidal flat below the present salt marsh, and thus have large error ranges, and a higher average elevation (Gehrels et al., 2012). Within these two points, SLIPs from this study are taken directly from the inferred transgressive surface and it is likely that deposition of the fine sediment, with the transgressing tide over the flooding surface, occurred at the higher end of the fossil's range. In this case the SLIPs could be slightly higher. Such small variation in elevation, which is smaller than the error range of the SLIP, cannot ever be fully resolved.

The observations at Little Swanport are consistent with the GHIA model predictions (Lambeck and Nakada, 1990), showing no evidence of a mid-Holocene highstand. Gehrels et al. (2012), show that sea level was still below present levels at around 650 cal yr BP. The SLIPs from South Neck Beach show more variation, and suggest sea level could have reached present levels by 2725 yr BP. However, those SLIPs were not focussed on sea level reconstruction and so are only proxy SLIPs, or “basal” as defined by Engelhart et al., (2011). Precise indicators are needed to resolve this, but by extending between the error limits of SLIPs between 5300 and 4000 cal yr BP, a very slow rate of SLR is evident until 650 cal yr BP.

4.4.3 Geochronological estuarine intertidal evolution

The processes of sea level transgression and estuarine infill can be identified in the Little Swanport sedimentary record from stratigraphic features evident in the cores. They show good similarity to the features identified by Sloss et al. (2004; 2006a, b; 2010) for the inference of processes and the timing of sea level transgression in New South Wales estuaries. The correlations are used here to aid interpretation of the sedimentary record in Little Swanport, using the same division of palaeo-geomorphic stages. However in this study a phase of continuous intertidal infill that accounts for the absence of a sea level highstand is added in place of the regressive phase described by Sloss et al. (2004; 2006a, b; 2010). Finer scale similarities are evident between cores from the present study in Little Swanport and those of Gehrels et al. (2012), and with the South Neck Beach stratigraphy (Clark et al., 2011). For clarity, the correlations focus on TCLB-1.

4.4.3.1 Stage 1 Late Pleistocene Transgressive Surface

The Pleistocene surface in this study was inferred from the dense, pale clay that would be expected in low lying areas of the incised valley cut into Jurassic dolerite (DPIPWE, 2004). Green mottles on the surface of the clay are associated with salt water intrusion, and the onset of anaerobic conditions (Keary, 2003). It contained plant fragments but no marine indicators, such as molluscs or foraminifera, beyond those found on the surface.

The stratigraphic record contained within the Little Swanport cores from the present study shows a high degree of correlation with those of Sloss et al. for Lake Illawarra (2005), Lake Burrill (2006), and Lake Conjola (2010), in Southern New South Wales, at the base of the cores. In both areas, palaeo-geomorphic Stage 1 is indicated by a reduced mottled basal substrate of clay. At Little Swanport this consisted of a dense white, yellow to brown clay with green mottling.

4.4.3.2 Stage 2 Mid Holocene Transgressive Sand Sheet Facies 5TCLB (Facies 5 TCLB1 & TCLB2 & Facies 3 MFLB1)

Lying unconformably over the late Pleistocene transgressive surface at Luttrells Bay and Lake Illawarra, New South Wales is a sand rich mud with rip up clasts of coarser material. Both contained broken and intact estuarine and marine molluscs indicative of transgressive processes. These processes occurred basin wide in Lake Illawarra. Evidence of these processes in cores from Luttrells Bay, and Duck Island in the upper estuary, suggest a similar geographic extent in Little Swanport as indicated by the predominance of fine sand with shell hash. A combination of low intertidal foraminifera and the occasional salt marsh species indicated that the palaeoenvironment of deposition was close to mean sea level in the micro tidal elevation range. Shallow subtidal bivalves were mixed with mid intertidal surface dwelling gastropods, near the base of facies 5 in TCLB-1 and TCLB-2 and facies 4 in MFLB. The mixed assemblage at similar points indicates that at the time, the rate of sea level rise exceeded sediment availability, causing an increase in water depth.

The presence of crystalline quartz fragment and other minerals such as biotite and feldspar indicated the sediment supply contained considerable terrestrial sediment. The age of the sand sheet ranges from 7362 cal yr BP, when sea level was still rising rapidly, and pinches out some time before 5057 cal yr BP (see Figure 4-11), when sea level rise had all but ceased.

More locally, finer scale correlations from the macrofossil record are evident between Luttrells Bay and the South Neck Beach palaeo-estuary. At 1.05 m depth relative to present MSL in TCLB-1, a radiocarbon date of 3875 ± 45 cal. yr BP was obtained from the point in the core where molluscs became abundant. Foraminifera were rare and dominated by low intertidal species. The depth and date correlates well with that of Clark et al. (2011) who also found a shell rich layer at a depth of -1.18 m below mean sea level, dated at 4218 – 3970 cal. yr BP, at South Neck Beach.

In the palaeo-estuary the layer consisted of fragments of the oyster, *Ostrea angassi,* whose habitat is sandy to muddy estuaries. The highly dense but fragmented layer had clearly been transported some distance into the estuary four millennia ago. Possible transport mechanisms include tidal currents, regressive beach processes, or tsunami (Clark et al., 2011), with good evidence for the latter. However, placed in the context of the present study, there is also circumstantial evidence that the fragmented *O. angassi* was transported and deposited by transgressive processes into the palaeo-estuary, as at Little Swanport, making its presence consistent with the predicted sea level rise from the glacio-hydro-isostatic adjustment model (see red dashed line in Figure 4-11).

The inferred processes from both locations are consistent with an increased rate of relative sea level rise after a period of stability between 6000 and 5000 yr BP indicated by the GHIA model (Lambeck and Nakada, 1990), although they do occur at a lower elevation than the model predicts. Such an increase in the magnitude of sea level rise could erode deposits in the intertidal zone, and transport them to the higher intertidal elevation. The processes inferred are of estuarine infill, as the sediment is composed of reworked estuarine material, indicated by the macrofossil habitat range, and more open conditions inferred by the higher content and larger grain size, compared to the overlying facies. These sedimentological and ecological features suggest the onset of a change in magnitude of sea level rise which is consistent with that indicated by the GHIA model.

4.4.3.3 Stage 3 Intermediate Infill (Facies 4 & 3 TCLB-1, TCLB-2 and MFLB2)

The period 4000 to 2500 yr BP is one of continued sea level rise according to the GHIA model but there are insufficient precise SLIPs to confirm this rise during that time. Nonetheless, continued sea level rise was supported in the present study by a slight coarsening upwards in facies 4, although all samples are still in the fine sediment range. Facies 4 in TCLB-1 and its correlative facies 4 in TCLB-2 and 2 in MFLB do however, all contain a high proportion of well sorted shell hash which could be considered coarse grained in an environment where fine sediment is dominant. Based on chrono-stratigraphy from Clark et al. (2011), it is likely that their SLIPs found at -0.85 m at 2776 cal yr BP and -0.55 m relative to PMSL at 1307 cal yr BP, correlate with facies 4 and 3 respectively, in TCLB-1.

Facies 3, TCLB-1 and correlative facies 3, TCLB-2 and 2, MFLB-1, contain a lower proportion of shell hash than facies 4. The fossil molluscs are predominantly intact, low intertidal mudflat species. The dominance of *K. scalarina*, which is a low intertidal bivalve that lives in sandy mud, and the gastropods *A. porphorea* and *V. australis,* that live on the surface within intertidal seagrasses, indicates that the macrofossils are in their life position and not transported. If this were otherwise, they would be broken and sorted. The smaller size and volume of shell hash indicates a fining up of transported material despite the absence of any significant change in mineral grain size which would be expected of a typical fining up sediment infill sequence. However, according to Reading (1986), a fining upward sequence may not occur when the sediment supply to the region is high in clay and silt, as is clearly the case in Luttrells Bay. It is also likely that the bedrock constrained mouth of the estuary has limited the transport and deposition of coarser marine-sourced sands into the estuary.

The possibility of the shell hash rich facies being a channel lag deposit can be discounted because of its presence in all cores collected in Luttrells Bay. Collectively these indications suggest that sea levels were still rising after 4000 yr BP because material is continuing to be transported into the estuary and ultimately the Bay and the rate of sea level rise was greater than the availability of sediment in Little Swanport. The dominance of intertidal mollusc and foraminifera species indicates that infilling has been continuous, and its elevation and time constraint indicates that MSL was still below present levels.

4.4.3.4 Stage 4 Continued Intertidal Infill

The GIA model reported by Gehrels et al., (2012) predicts a drop of 0.25 m in relative MSL, to around its pre-anthropogenic induced sea level rise position, from 2500 yr BP. It is important to consider the magnitude of this change in terms of detection using palaeoreconstruction techniques. Considering potential error in the SLIPs available for this period, the results of the present study agree with the GHIA model, as far as there is no evidence of a mid-Holocene highstand of the magnitude experienced north of Tasmania. None of the transects characterising the marsh upland border, provided any geomorphological evidence of a raised shoreline, and all samples scanned at those elevations were devoid of molluscs or foraminifera. It is possible however, that any raised estuarine muds have been eroded and are thus no longer evident. Material from the surrounding terrestrial slopes may also have been transported and deposited to mask any landforms resulting from higher sea level.

Sea level in southern Tasmania was estimated to be 0.55 m below its present elevation around 1200 yr BP (Clark et al., 2011) and 0.29 m below its present elevation around 500 yr BP (Gehrels, 2012). This equates to a calculated rate of sea level rise of 0.037 m/year. Facies 4 TCLB-1 and facies 1 MFCLB-2, both contain assemblages more representative of largely intact, estuarine mudflat species although the sediment was still composed of well sorted shell fragments. This indicates that the transport of material was still a dominant process thus, sea level transgression was still a forcing factor in the geomorphological evolution of the estuary, albeit very slowly.

The Duck Island cores did not provide a salt marsh facies significantly deeper than at Luttrells Bay. It may be inferred from the almost featureless facies 3 and 4 however, that it is a bay head delta composed of fine to medium sand transported inland with sea level transgression and terrestrial sands from fluvial inputs.

4.4.3.5 Stage 5 Present (Facies 2 & 1 TCLB-1, C2LB Salt Marsh)

The transition from intertidal flats to pioneering salt marsh is evident from the gradational contacts between facies 3 TCLB-1 and C2LB. Both consisted of inorganic fine sandy silts and low intertidal foraminifera, to organic fine sandy silts, containing a mixture of salt marsh and low intertidal foraminifera in their respective facies 2. Rootlets were present throughout facies 2, but not at a density which would suggest that marsh species had fully occupied it at the surface. Facies 1 in both cores consisted of organic fine sandy silts, a dense root mat and abundant salt marsh foraminifera; intertidal species were absent. Only a minor amount of deterioration in the root system of the *Sarcocornia* spp. was evident, suggesting that marsh evolution has been relatively recent. Gehrels et al., (2012) proposed an age of 650 cal. yrs BP, at the base of the salt marsh soils in Watch House Bay (Table 4.1), at around the same depth as the base of facies 1 in Luttrells Bay, suggesting that the fringing marshes are of the same age.

The recent development of the marshes is shown by < 0.5 m of marsh facies in cores (Figure 4.11) and significant because their elevation range is tightly constrained to a tidal inundation period which gives insight into marsh evolution processes. Mature, brackish, low saltmarsh in Little Swanport occur within an elevation range of 11 to 18 cm above MSL at the seaward edge (see Section 3.4.2). Notably, facies 1 occurs to 50 cm depth in TCLB-1, a core taken at an elevation of 43 cm above PMSL, indicating that its development commenced when paleao MSL reached around +7 cm. This correlates well with the present prograding salt marsh shoreline.

Although the overlying salt marsh facies were deeper on Duck Island at 0.6 to 0.7 m compared to the shallower fringing marshes, their base still occurred at around 0.1 m below present MSL, suggesting that it reached an elevation suitable for salt marsh establishment at a similar time. The shorelines of Duck Island exhibited net progradation so is likely to be receiving adequate sediment to keep pace with present rates of sea level rise and suggests that the estuary is in the latter stages of infilling which is further supported by the mature geomorphology of the fringing marshes.

The palaeo-geomorphic stages and environments described above are consistent with the wave-dominated estuarine infill model of Dalrymple et al. (1992). The inductive model has been a useful aid for predicting suitable locations for coring in the estuarine setting and the palaeo-environmental interpretation of the sediment stratigraphy in the present study. The regional chronostratigraphic framework recommended by Sloss et al. (2010) is similarly useful.

4.5 Conclusion

Conformably underlying shallow salt marsh, in a quiescent, micro tidal estuarine setting, is a sedimentary record of estuarine evolution with sea level transgression which was reconstructed using precise coring and elevation surveying methods. The site selected in this study has been shown to be suitable because when remote from fluvial inputs, marine influences are the dominant forcing factor. The results are consistent with the palaeo-environmental interpretation of the estuary enabled by the inductive model of Dalrymple et al., (1992).

Precise sea level indices were obtained by retrieving dateable material directly from the inferred transgressive surface. These are base of basal samples, and thus suitable for inclusion in glacio-isostatic adjustment models (Engelhart et al., 2011a). Their locations on the inferred late Pleistocene transgressive surface, with depths of mostly 1.5 m (Figure 4.11), were obtained in a manner designed to reduce the error margin, other than for compaction of the underlying clay, which is likely to be negligible as clays expand upon saturation.

The present study has confirmed the feasibility of using the chronostratigraphic framework of estuarine evolution with sea level change recommended by Sloss et al., (2010), for south eastern Australian studies. The stratigraphic correlations across sites, particularly for macro fossils of intertidal molluscs, has enabled the inference of the processes of deposition from sea level transgression to sedimentary intertidal infill which continues to present in Tasmania. The species turnover from the transgressive sand sheet to intertidal infill facies in Tasmania, did not indicate any increase or reduction in marine influence which was interpreted to have occurred with back barrier development in New South Wales, and the sea level fall from highstand (Sloss et al., 2004, 2006a, b; 2010).

The apparent local variation in sea level history for both Tasmania and New South Wales are in good agreement with the predictions generated by the glacio-hydro -isostatic adjustment model of Lambeck and Nakada (1990). For south eastern Tasmania sea levels rose relatively rapidly until around 7000 years BP and slowed after this time till around 6000 years BP. The rate of rise is consistent with the model predictions, however, sea level was still around 0.5 m below present MSL at this time, rather than the predicted 0.25 m above present MSL. Further, there is no evidence of a mid-Holocene sea level highstand for Tasmania although this needs further investigation to confirm and fill the late Holocene record gap for Tasmania. The intertidal facies underlying the salt marsh facies could provide sea level index points, if a modern analogue of this environment can be constrained to a narrow elevation range, and potentially included in a transfer function training set.

Chapter 5 Late Holocene sea level stability in Tasmania, south eastern Australia: A full intertidal transfer function broadly indicates the nature of Late Holocene sea level change

5.1 Introduction

Salt marsh elevation and foraminifera transfer functions are commonly employed to quantify late Holocene sea level change. The method allows for replication and cross-comparison of records (Horton and Edwards, 2006; Wright et al., 2011; Kemp and Telford, 2015). Foraminifera-based transfer functions (FBTF) have enabled local to global comparison of accurate local sea level index points (SLIPs) between sites (Horton and Edwards, 2006; Shennan et al., 2006; Engelhart and Horton 2012; Gehrels et al., 2012; Kemp et al., 2013). Such comparisons are vital in modelling Earth's rheological responses to the changing mass balance of ice to ocean volume through the last marine transgression and to anthropogenically-induced climate change (Shennan and Horton, 2002; Kemp et al., 2009, 2014; Gehrels et al., 2012; Barnett et al., 2016).

Salt marsh foraminifera are few but cosmopolitan. Their elevational zonation is well known for the Atlantic coast, north eastern America (Scott and Medioli, 1978; Gehrels, 1994; Gehrels et al., 1996; Horton et al., 2009; Leorri and Martin, 2009; Wright et al., 2011; Kemp et al., 2012, 2013; Long et al., 2012;) the British Isles (Horton, 1999; Horton et al., 1999; Gehrels et al, 2001; Horton and Edwards, 2006) Australasia (Southall et al., 2006; Gehrels et al., 2008; Callard et al., 2011), the latter in south eastern Tasmania. Data are also available for other locations, such as Canada (Barnett et al., 2016) the Adriatic coast of Croatia (Shaw et al., 2016) and the southern coastline of South Africa (Strachan et al., 2016).

In undertaking analyses with FBTF, species abundances are recorded for samples of known elevations. Species optima are expressed as mean elevation and tolerance limits are approximated by the standard error. These values are used in a transfer function to indicate the elevations at the time of deposition from fossil foraminifera. The overall accuracy of the transfer function is expressed as an error margin by regression parameters, including the root mean squared error of prediction (RMSEP) and r^2 bootstrapped (r^2_{boot}), which indicates precision. Micro tidal ranges provide high precision FBTF, with RMSEP ranging from ± 0.05 to 0.15 m (Scott and Medioli 1980; Southall et al., 2006; Callard et al., 2011; Barnett et al., 2016; Shaw et al., 2016). The longer elevation gradients of macro tidal ranges, yield a smaller precision of around ± 0.26 – 0.23 m for vegetated tidal ranges and ± 0.38 – 0.31 when the gradient is extended to lowest astronomical tide (Woodroffe, 2006).

Problems in using FBTF to generate SLIPs include potential spatial autocorrelation associated with regular, systematic sampling of the compact salt marsh environmental gradient (Whysong and Miller, 1987; Telford and Birks, 2005). Wright et al. (2011) demonstrated that closely spaced samples are less dissimilar in ordination space than samples not closely spaced but of the same elevation ranges, so typical transfer function weighted average regressions may overestimate precision. Truncated elevation gradient sampling can decrease error ranges around species optima (Woodroffe, 2009). This problem can be resolved by sampling the entire gradient, rather than just sampling the range found in the fossil assemblage (Wright et al., 2011).

Other sampling issues include the risk of missing analogues and spurious correlations between the training set and fossil assemblages caused by spatial and seasonal variability of species within transects (Horton and Edwards, 2006; Wright et al., 2011). These risks can be minimised by multiple transect sampling within sites, (Kemp et al., 2011) or regional scale sampling (Horton and Edwards, 2006) for the modern analogue. Inaccuracies related to seasonal variation in live population abundances can be mitigated by sampling in winter only, and using dead foraminifera from surface sediments (Horton and Edwards, 2006: Callard et al., 2011; Wright et al., 2011; Kemp et al., 2015).

Apart from inundation period, other environmental factors, such as salinity, can affect species turnover along an elevation gradient (De Rijk, 1995; De Rijk and Troelstra, 1997; Abu- Zeid and Buntan, 2013) and vary by season (Horton and Murray, 2007), making it important to understand their influences before using foraminifera as indicators of past sea levels (Hortonet al., 1999; Horton and Murray, 2007; Strachan et al., 2016).

Wright et al. (2011) make two recommendations to improve sampling standards. They are, extending the training set to include the whole species response curve, and including contemporary samples from other sites to avoid missing analogues in the fossil record. Recommended improvements to statistical handling include employment of ordination techniques to describe the similarity between the modern analogue and the fossil assemblages. If the FBTF development process is handled explicitly, the transfer function can be a tool for answering specific research questions (Wright et al., 2011). In the present case, the transfer function approach was applied to fill gaps in local sea level records.

Salt marsh facies have been retrieved at greater than 2 m depth, with ages of around 2000 yr BP, providing precise indicators of late Holocene sea levels (Engelhart et al., 2011a). However, Tasmanian salt marsh facies have not been observed deeper than 50 - 60 cm (Gehrels et al., 2012; Morrison and Ellison, 2017). The deepest precise SLIPs from salt marsh facies in Luttrells Bay yielded a ^{14}C age of 650 years BP (Gehrels et al., 2012). Other high precision sea level index points were developed by dating intertidal macro fossils, systematically located on the inferred late Pleistocene flooding surface, shown to increase in elevation contiguously with the present valley slope (see section 4. 4). Former mean sea level position was obtained from 7363 ± 71 to 4665 ± 126 yr BP, with accuracies ranging from ± 20 to ± 40 cm (figure 4-1). However, samples could not be located higher along the slope gradient in any part of the bay to obtain later Holocene SLIPs, and so there is a gap in precise SLIPs for the Late Holocene.

Figure 5-1 shows two SLIPs derived from a palaeo – tsunami study (Clark et al., 2011), that indicate a sea level rise from around 4000 yr BP to present levels at 2700 yr BP. Although these are not precise indicators the trend is consistent with the GIA predictions as indicated by Gehrels et al. (2012), but the rising trend is then followed by a sea level fall to - 0.5 m at 1200 yr BP. Earlier studies on Tasmanian sea level history similarly suggest a minor sea level highstand and fall at around 2330 yr BP (Haworth et al., 2002). The lack of precision, the few SLIPs and previous studies suggest 2 options: 1) there was a late Holocene sea level highstand consisting of a minor rise and fall, or 2), there was a slow gradual rise from 0.5 m to 0.25 m at 650 yr BP. High accuracy SLIPs are needed to test these options. Thus, the suitability of a FBTF that extends from salt marsh to the low intertidal zone is worthy of investigating to create SLIPs for the late Holocene.

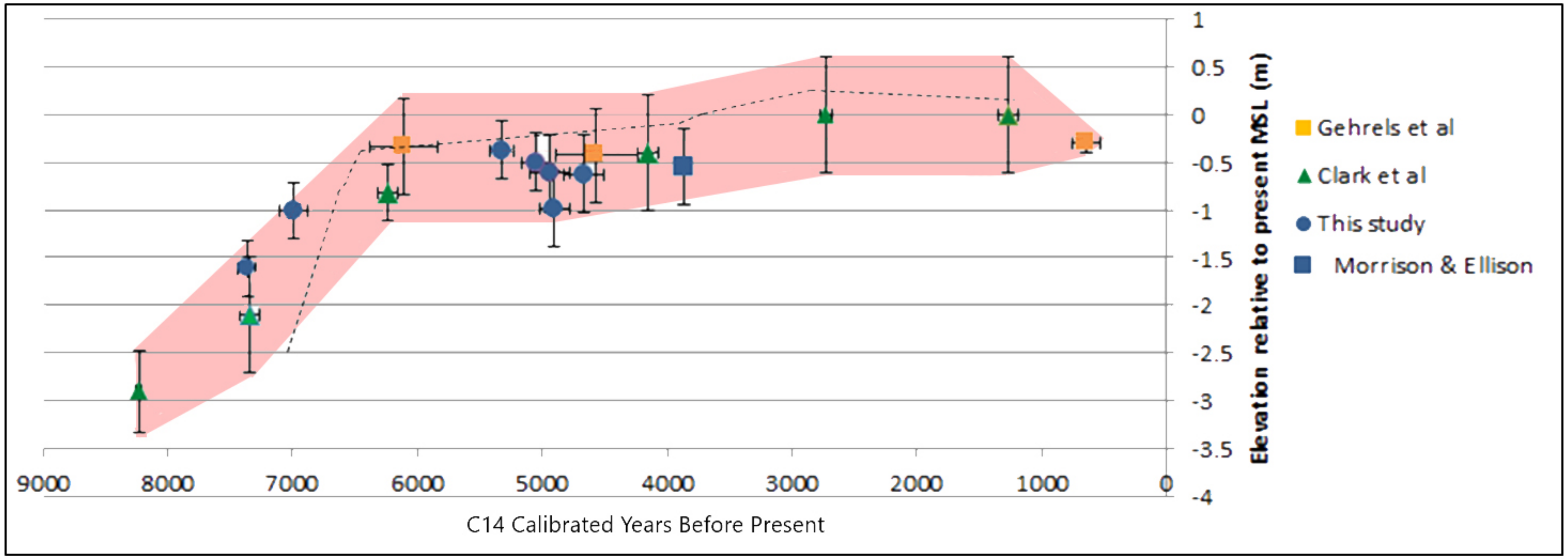


Figure 5-1 Sea level curve for south eastern Australia showing a rapid rate of sea level rise from 8000 to 6000 years BP followed by a period of relatively stability from 6000 to 3700 years BP. After 3700 years BP there is a gap in high precision sea level index points so the nature of change from 3700 to 650 years BP is presently unclear (see Chapter 4). The black dashed line shows the glacio-hydro isostatic model predictions stated in Gehrels et al. (2012), and the pink shading indicates the potential vertical error which are explained in section Chapter 4.4.2.

In any tidal range, low intertidal modern analogues are not as tightly constrained to elevation, because salinity and other strong limiting factors co-vary less with elevation below MHT level than above MHT (Woodroffe, 2009; Hayward et al., 2010). In a macro tidal mangrove setting, this effect adjusts the elevation optimum of some species upwards, leading to a higher RMSEP of the FBTF (Woodroffe, 2009). The plausibility of the full intertidal transfer function is further reduced because subtidal facies cannot be discerned from low intertidal facies in sediment cores, using either biological or lithological features. However, Woodroffe (2009) recommends that a full intertidal FBTF be tested in a temperate micro tidal setting, particularly if the low intertidal zone can be discerned from the shallow subtidal zone.

The rationale for the present study is that previously reported sediment cores from Luttrells Bay record intact mollusc assemblages that occur commonly in the modern intertidal habitats. The mollusc assemblages and their intact nature, including fully articulated bivalves, indicates that molluscs are in their life position, and are useful as indicators of past environments. According to Tomašových and Kidwell (2009), molluscs are predominantly distributed according to environmental gradients, rather than dispersal limitations. In Luttrells Bay, these molluscs occur across the tidal flats, to just below salt marsh seaward edges. Marsh edges in different tidal ranges in Tasmania are limited to variable inundation periods but in Luttrells Bay occur at around 33 % inundation period, equating to around mean high tide position, so offer good elevation control (see section 3.4.2).

The aim of the present study is to extend the range of an FBTF, by testing the hypothesis that sufficient species turnover occurs across the whole intertidal zone, such that a transfer function will provide elevation constraint smaller than the likely sea level change. The primary objective is to calibrate the FBTF with a core, and identify suitable material to develop SLIPs that could fill the gap in the current sea level curve.

To achieve these objectives:

1) transects were placed along morphologically distinct seaward edge types in Luttrells Bay to capture as much environmental diversity as possible.

2) the relative influences on foraminiferal distribution of elevation and physical and chemical factors were determined.

3) transfer function performance of different elevation gradient lengths was tested, and

4) the most ecologically plausible transfer function was calibrated with a sediment core and evaluated for its suitability to broadly describe sea level history during the target time span.

5.2 Methods

5.2.1 Study Site

Two distinctly different geomorphological seaward edge types were evident in Luttrells Bay. These types reflected prevailing wind speeds and direction. Along the windward side of Luttrells Bay, experiencing stronger, more persistent winter wind speeds the average marsh edge elevation was 11 cm ± 4 cm (see section 3.4). Slumping edges were probably due to long periods of tidal inundation leading to poor drainage and a perennially saturated low marsh, (see Figure 5-3a and b). *Samolus repens*, *Sarcocornia blackiana* and *Sarcocornia quinqueflora* occurred in a mosaic distribution with infrequent *Tecticornia arbuscula*. By contrast, the leeward side of the marsh experienced lower wind speeds. It had a higher mean elevation of 18 cm ± 4 cm and the seaward edge was drier and did not inundate during normal high tides, (see Figure 5-3c). It displayed a cliffed neck and cleft formation, typical of cyclical phases of erosion and progradation driven by process feedback mechanisms described by Schwimmer and Pizzuto

(2000). Vegetation distribution across this section of low marsh was uniform, with a narrow zone dominated by *S. repens* along the seaward edge. *Sarcocornia blackiana* dominated the rest of the low marsh. These biogeomorphic distinctions are consistent with the well understood role of wind generated sheer stresses in sediment erosion and deposition (Fagherazzi and Wyberg 2009; Hunt et al, 2015). Because the biogeomorphic variants had been previously shown to be strongly related to the environmental variations of interest (see Section 3.6), they were sampled and included in the training set to represent modern analogues of variation, normal in marsh evolution. Missing analogues for core calibration can be avoided in this way (Gehrels et al., 2001; Southall et al., 2006; Szornick, 2006).

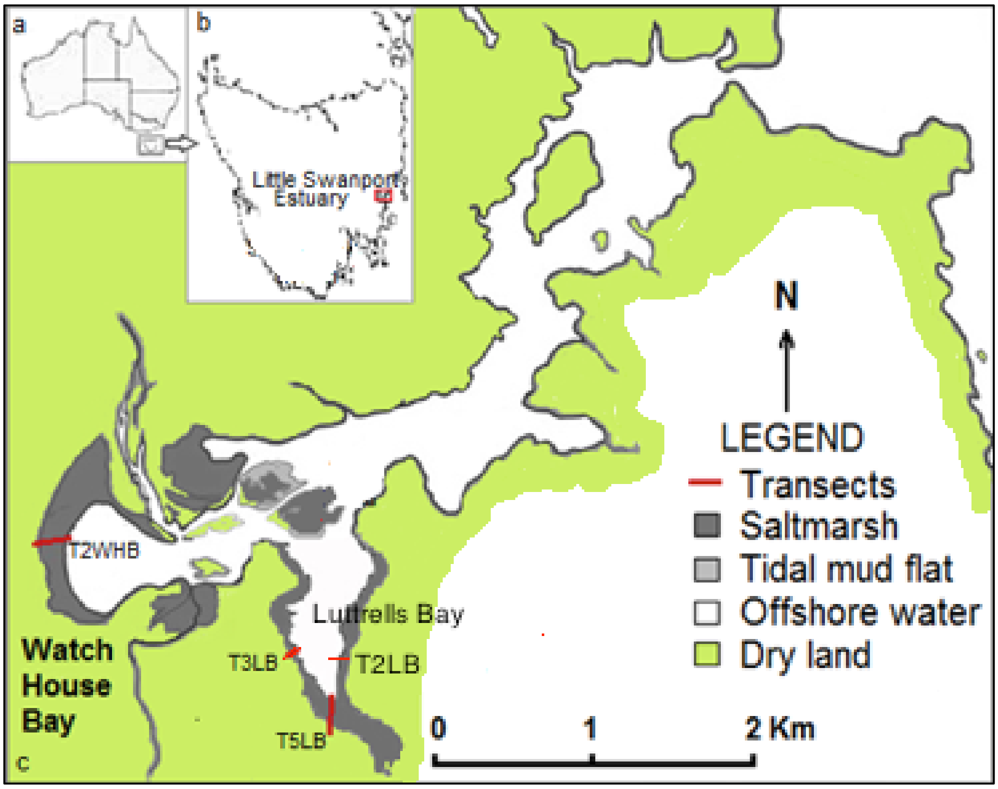


Figure 5-2 Luttrells Bay within the Little Swanport Estuary, south eastern Tasmania showing sampling locations for this study in Luttrells Bay, collected from previously described transects (Section 3.4.1), and T2WHB where the samples for Callard et al. (2011) transfer function study were collected. Sixteen salt marsh only samples from T3LB were included in the latter study.

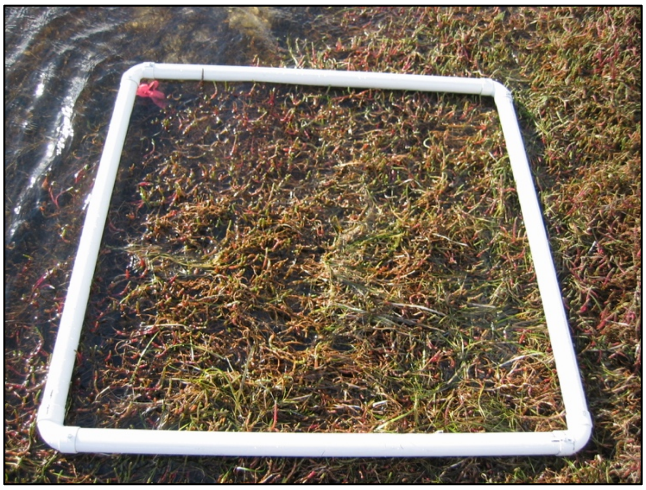


Figure 5-3a The seaward edge of transect 2 on the windward side of Luttrells Bay that had an average elevation of 11cm ± 4cm. It was inundated by mean high tides.

Figure 5-3b The eroding marsh seaward edge of transect 5 showing sampling locations at elevations that were inundated by daily high tides. The vegetated elevations were not inundated daily.

Figure 5-3c The seaward edge on the leeward side of Luttrells Bay displaying a neck and cleft formation and had an average elevation of 18cm ±4 cm. Three vegetation zones are evident from the very low marsh zone 1 edge dominated by a mosaic distribution of *S. repens* and *S. quinqueflora* (green) and zone 2 low marsh dominated by *S. blackiana*. The mid marsh zone 3 dominated by *J. kraussi* and zone 4 tussock grasses. Transect 3 was located on this side of the marsh and the seaward edge and tidal flat were sampled.

5.2.2 Elevation control, surface sampling and coring

A 10 x 10 x 1 cm surface sample was collected along Transect 5 at Luttrells Bay (T5LB) at every 2 cm fall in elevation, from the marsh upland border to the lowest obtainable tidal flat s during spring low tide for foraminiferal and sediment analysis. The seaward edge and tidal mud flat was sampled from Transect 2 (T2LB). Additional upper tidal flat samples were taken from Transect 3 (T3LB), see Figure 5-4. All elevations were obtained by a closed level run using a NA270 Level from a benchmark of known height, tied to mean sea level according to Australian Height Datum (see Section 3.2.3).

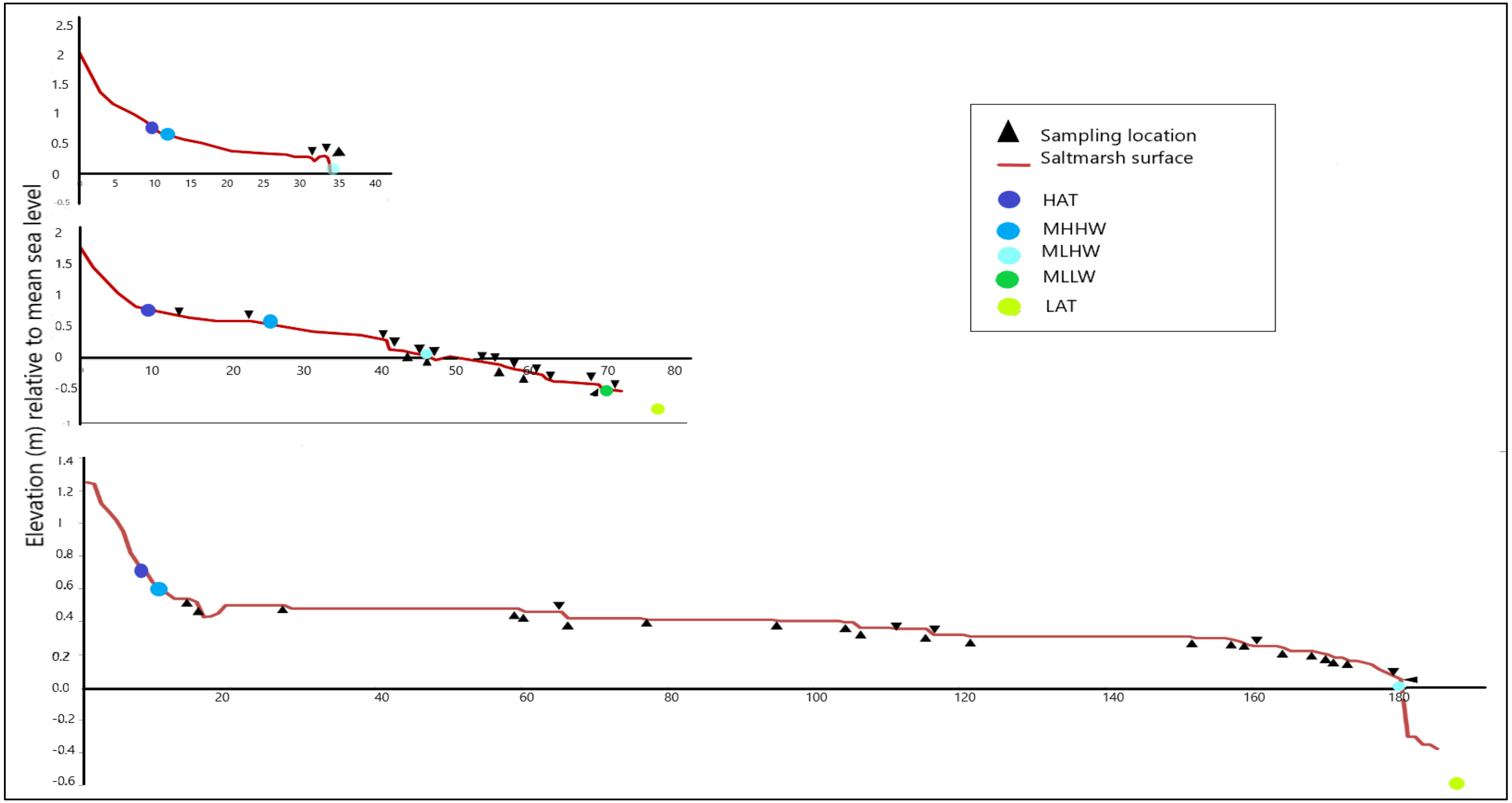


Figure 5-4 Sediment and Foraminifera sampling locations, along T2LB, T3LB and T5LB with tidal planes indicated for each transect**.**

On transect 5, also surveyed for elevation (see section 4.4), one core was extracted from 0.41 m above MSL, using a 5 cm diameter, 1 meter barrel hand operated Hiller Piston corer. The core was located at a distance of 1 meter from TCLB2 described in Chapter 4, but at the same height by being just offset from T5LB. The core was extruded into halved PVC pipe and transported to the laboratory.

5.2.3 Sediment analysis

Sediment pH was measured on the 10 x 10 x 1 cm sample within 24 hours of collection, with dampened pH paper test strips (with 0.2 pH increments) that were pressed into the sediment and left until there was no further colour change. Three replicates were taken and the results averaged. Sub samples of 5 ml were taken, weighed, dried at 60° C for 48 hours and reweighed for percent moisture content. To measure sediment salinity the same samples were then pulverised and added to 40 ml of distilled water, shaken and left for 24 hours. A further 100 ml was added to provide enough solution for the sediment to settle, and to ensure that the remaining volume was enough to cover the salinity probe. The hydrometer method was used to quantify the relative proportion of sand, silt and clay after Gee and Bauder (1986), using a minimum of 35 gm of sediment. Five ml of sample was taken for organic content analysis. Organic carbon content was determined by the loss on ignition method for samples with high organic content after Heiri et al. (2001).

5.2.4 Foraminiferal taxonomy and analysis

The stained surface subsamples were washed through sieve sizes 500, 125 and 63 *um*. The 125 and 63 micron fractions were retained and placed in a wet splitter for further subsampling. Due to the very dense *Sarcocornia* root mat, sieving yielded relatively little sediment for some samples. The live and dead population of Foraminifera were counted after Gehrels, (2002), to 150 individuals if abundant, and a minimum of 100 if less abundant. A minimum of 50 individuals were picked and mounted on gummed micro- paleontological slides. Population census data are expressed in percent relative abundance, calculated as the number of individuals of a species in a sample, divided by the total number of individuals of all species in a sample.

The foraminiferal data set includes species that occurred with 2% relative abundance or greater. Species that occur with less than 2% relative abundance are usually removed from the training set because they are allochthonous, and not representative of the endemic micro flora (Horton and Edwards, 2006).

Two taxonomic guides were used to identify species. Hayward et al. (1999) describe the distinguishing features of recent New Zealand shallow water benthic foraminifera, and includes thorough descriptions of species ecological range, global and local distributions, and remarks on classification issues. Yassini and Jones (1995) provide descriptions for Australian estuarine and shelf environments for south eastern Australia. Distinguishing features of species are described, including shape, colour and general teste form (see Appendix 5.1).

5.2.5 Data handling

Multivariate analyses were conducted in open source R, version 3.2.3, and the transfer function was calculated in C2 using default options (Juggins, 2007). The transfer function was developed using the steps outlined in Figure 2.7, commencing with ordination and classification followed by direct gradient analysis and weighted averaging core calibration.

5.2.5.1 Ordination and classification

Ordination techniques were employed to explore the species data and identify similar samples by groupings as recommended by Wright et al., (2011). Agglomerative Hierarchical Clustering by Ward's Method was used to identify the number of groups for samples. Ward's method avoids chaining in dendrograms by fusing on the basis of mean Euclidean distance value, rather than on the basis of mean attribute values. The number of groups was assessed by scree plot examination. To visualise the groups, non-metric multidimensional scaling ordination was employed because it produces better outcomes than the factor analysis family with matrices with many zeros (Minchin, 1987).

5.2.5.2 Direct gradient analysis

Detrended correspondence analysis was conducted first on the whole species data set to identify gradient length (ter Braak and Verdonshcot, 1995). For studies aimed at theory generation, DCA is a controversial technique, criticised for its artificiality because it imposes assumptions about the distribution of sample units, and species in environmental space. For this reason McCune and Grace (2002), find no reason to recommend it, but ter Braak and Limneaur (1995) state that for the purpose of defining gradient length as a function of the standard deviation (SD) of species responses, it is useful for quantitative palaeoecology.

Canonical correspondence analysis (CCA), was conducted on the full species and environmental data set, and the ANOVA function in R, used to test for statistical significance of the model. CCA is an eigenvector ordination technique that extracts synthetic gradients (ordination axes), from measured environmental gradients and maximises the separation of species along them (ter Braak and Limneaur 1995). It is applied in quantitative palaeoecology for statistical testing of variation in multiple species responses to environmental variables (Borcard et al., 1992; ter Braak and Limneaur 1995), where variance is a measure of species dispersal. It was used to determine if elevation is the primary determinant of composition of foraminifera, warranting the development of a FBTF (ter Braak and Limneaur 1995). It was not used for causal interpretation of species patterns (Borcard et al., 1992; McCune and Grace, 2002). Partial CCA is also used to determine the interactions between environmental variables in explaining variation in species composition.

The analysis occurred in three stages. Firstly CCA was conducted on the matrix of environmental variables: elevation, sediment salinity, pH, % sand, silt and clay content, and, organic and moisture content. Model simplification was conducted by ecological reasoning and stepwise forward deletion of variables that had no significant effect on the species data, with a cut off value of p =/< 0.05. The variance inflation factor (VIF) was examined for each remaining variable, and those with a factor of over 10 were removed from the data set because this indicates too great a co-linearity with other variables for model integrity. In the second stage CCA was conducted on the reduced model to determine the total variance in the species data (total inertia).

Stage 3 employed Partial CCA to separate out the species variance by an individual environmental variable, from the contribution of the covariables, achieved by making the secondary or later axes orthogonal (ter Braak and Verdonshcot 1995; Oksanen, 2015).

5.2.5.3 Transfer function, weighted average regressions.

Five versions of weighted averages methods for regression and calibration were applied to find the best performing model of species and elevation, as assessed by the prediction error (RMSEP), and the correlation coefficient (r). Based on the unimodal response, the weighted average methods examine individual species optima and tolerances, which is important because it allows calibration of ‘non-analogue’ situations, outliers and species absences (Birks et al., 2010).

Model selection was based on the principle of parsimony in that it should provide the lowest root mean square error of prediction (RMSEP, or the vertical error), and lowest maximum bias (the difference between the mean observed and predicted values), but maximise the covariance between species weighted averages and environmental variables, using the least components (Massey et al., 2006; Birks et al., 2010). The coefficient of determination r^2, was assessed using cross validation techniques of bootstrapping with 1000 iterations and conducted on all models except Partial Least Square Weighted averaging (PLS-WA), which employed Jacknifing in C2. Weighted averaging core calibration was conducted using the modelled response from the regression to infer environmental conditions from the composition of fossil foraminifera assemblages. Stratigraphically constrained cluster analysis (CONISS) was run on the fossil data to evaluate the reconstruction and interpretation of palaeoecological phases based on facies lithostratigraphic and micro and macro biostratigraphic features.

5.3 Results

5.3.1 Species and environment

5.3.1.1 Intertidal foraminifera

A total of 18 species were identified in the full data set from the Little Swanport Estuary. Samples from un-vegetated intertidal zones in the upper estuary, outside of Luttrells Bay were dominated by calcareous species including *Elphidium crispum* (> 80% relative abundance). Other *Elphidium* spp., *Miliammina fusca* and *Miliammina oblonga* occurred with less than 5% relative abundance. These samples were not included in the species data set due to their unlikely presence in Luttrells Bay, indicated by baseline investigations (Morrison, 2006), and the FBTF developed by Callard et al. (2011).

All samples across the intertidal gradient in Luttrells Bay were dominated by agglutinated species but there was an increase in calcareous species below mean sea level on the intertidal flats. Calcareous species were more often greater than 60% alive, and *Quinqueloculina seminula* was > 80% alive on the tidal flats but only 46% alive on the salt marsh. However, it was difficult to discrimate between live and dead for darker species, such as *M. fusca*, and coarsely cemented agglutinated species with high proportions of dark mica sediment particles, such as *Ammobaculites exigus* and *Ammobaculites agglutinans.*

5.3.1.2 Classification

Cluster analysis showed 5 groupings (cophenetic correlation coefficient r = 0.677), (see Figure 5-5*)*.

Figure 5-5 Cluster analysis on foraminifera samples showing 4 groupings (cophenetic correlation coefficient r = 0.677). The largest grouping (a) consists of samples from the mid and upper marsh. The next largest group (b) consists mainly of samples from the very low marsh and seaward edge of marsh. Group c are mainly from the lowest mean low tide mark extending to the edge of the sea grass patches. Group d are samples are from the edge of the upper tidal flat and include sample 37 which was taken from the seaward edge of Transect 2.

The NMDS showed good correlation between the original calculated sample distances and the final sample configuration (R^2 = 0.893, stress = 0.17). The four groups evident by cluster analysis were also discernible in the NMDS biplot. The low intertidal mud flat groups (c is brown and d is purple) plot to the left on axis 1. The highest tidal flat samples 12, 13 and 14 in group d, are separated from samples 1 to 11 which are spread on axis 2, including the most distant, sample 37.

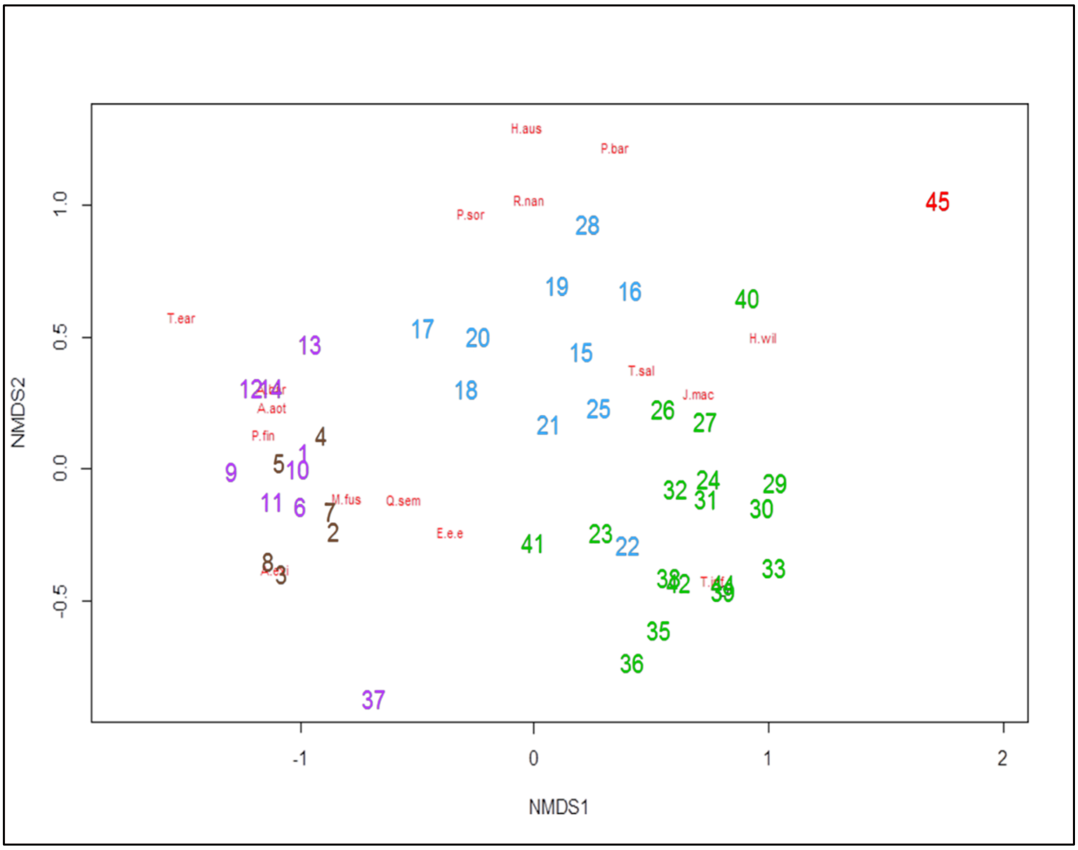


Figure 5-6 Non metric multi-dimensional scaling of the full species and sample data set (stress = 0.174) showing the distance between samples is greatest between tidal flat groups (group c is brown and d is purple), and salt marsh groups (group b is blue and a is green) along axis 1.

On the right of NMDS axis 1, the two salt marsh groupings are evident, and are gradually spaced along axis 1 and 2 (see Figure 5-6). Group b, (blue) is comprised of seaward edge marsh to low marsh samples, 15 – 25 (and 28), collected from the low end of the 3 transects, spanning an elevation range from just above MSL at +

0.03 m, which approximates MLHW at Luttrells Bay (+ 0.029 m), to + 0.19 m, equating to annual inundation period range of 38 - 25%. This elevation range section is dominated by *Miliammina fusca* and *Trochammina inflata*, with sub-dominant *Reophax nana* which is limited in occurrence to this grouping only, and has an average relative abundance of 24%. Calcareous species *Ammonia aoteana* and *Elphidium excavatum* were less abundant but occurred in most samples, along with calcareous species *Protoschista findens* and *Ammobaculites exigus*. The presence of both tidal flat and salt marsh dominants indicates that this elevation range is an ecotone between tidal flat and salt marsh environments.

The salt marsh group (a), consists of samples 23 - 44 at T5LB, and samples 40 and 41 from T2LB. Samples 23 to 33 range in elevation from + 0.21 m, up to + 0.52 m which is around MLHW to close to MHHW and are mostly faithful to a subgroup of group a. The inundation period is broad, from approximately 22% to 2% per annum. These mid to upper marsh samples are different to the preceding group because *M. fusca* is no longer present, but the assemblages are still dominated by *Trochammina inflata* and *Jadammina macrescens*. The latter is subdominant and increases in abundance across the elevation range from 2% to up to 28%. The subordinate species, *Haplophragmoides wilberti*, decreases with elevation from 18% relative abundance to 3%. This group also contains the minor species *Trochamminita salsa.*

Samples from 33 – 44 in group **a**, plot to the bottom right of the group, distanced from 23 – 32 along axis 2. These samples are located from elevations above + 0.57 m which approximates MHHW (0.60 m at Little Swanport), to +0.78 cm which approximates highest astronomical tide (+ 0.75 m). This elevation range experiences an annual tidal inundation period from 1.3 %, to < 0.01%. *Trochammina inflata* is faithful to this high marsh group and dominates all samples at greater than > 80% relative abundance, and often exceed 90 %. However, in the more elevated samples (43- 44), all individuals were dead with low individual counts of < 50.

Outliers include sample 37 which was obtained from the very seaward edge off Transect 2. Its site was morphologically unique, in that it exhibited micro slumping, poor drainage, low plant vigour with some live sea grass (see plate 5-3a). It was the only tidal flat sample in which *Ammobaculites agglutinans* dominated *M. fusca* (> 90%), and the individual count was high. It was retained in the data set because it represented a seaward edge type typical of marsh evolution (see 5.2.1). Sample 45, a singleton sub-group in **b**, was taken from 63 cm above MSL, which is above MHHW and is an extreme outlier sample consisting of 100% dead *Haplophragmoides wilberti* and very low individual count of 9. *Portatrochammina sorosa* and *Paratrochammina bartrami* occurred with less than 2% relative abundance, in the low marsh and tidal flat samples and in less than 5% of samples. They are not described elsewhere as saltmarsh or intertidal species (Hayward et al., 1999) and were considered allochthonous, so were removed from the training set following Horton and Edwards, (2006). Dead *Textularia earlandi,* associated with the smaller tidal flat group, occurred in all three samples but with relative abundance below 2% and so was removed.

5.3.2 Environmental variation in the training set (ordination)

The gradient lengths for DCA were greater than two standard deviations and so analysis progressed with direct gradient analysis by CCA, suitable to unimodal gradients (Table 5-1).

Table 5-1 DCA parameters of the full foraminifera and environment training set

	DCA 1	DCA2	DCA3	DCA4
Eigenvalues	0.777	0.2745	0.212	0.268
Gradient length (SD)	3.446	0.652	1.6039	1.2310

All variables in the full environmental matrix were significant at the p= < 0.001 level (see Table 5-2), except pH which was not significant at the p = < 0.5 level, and was removed from further analysis. Organic content was highly correlated with sediment salinity and had a variance inflation factor of 23.4, and was removed from the data set. The variance inflation factor for silt and clay content was also greater than 10 indicating high co-linearity, and was removed from the data set, which lowered the VIF of % sand content to less than 10. The VIF of all remaining significant variables was less than 10.

Analysis by CCA of the reduced model including sediment salinity, % sand content, moisture content and elevation showed that 51% of variance in the foraminifera data set was explained (Table 5-2). Variance partitioning by partial CCA showed that elevation explained the highest unique portion of the species variance at 6.4%. Salinity and moisture content contributed 3.9% and 3% respectively, and sand content explained the smallest portion of 2.7%. Therefore, the total unique contributions totalled 16% and the remaining 35% of the species variance is explained by the interactions between variables, which have different relationships in different parts of the gradient (see Figure 5-7). The analysis indicated that inundation period was the most important environmental variable in influencing foraminifera. The amount of variance that is explicable by the environmental data is moderate, but this is normal for CCA analyses. Interpretation focusses on relative, not absolute, amounts of variance (Borcard et al., 1992; Ter Braak and Limneaur, 1995).

Table 5-2 CCA for the reduced model including elevation, salinity, sand and moisture content, individual variables and variance partitioning parameters for the significant environmental variables of the reduced model. (SSE – salinity, sand content, and elevation, MSEa moisture content, salinity and elevation, MSEb moisture content, sand content and elevation).

CCA Model	**Total inertia**	**Constrained Inertia**	**Total variance %**	**p =**	**CCA1**	**CCA2**	**CCA3**	**CCA4**
Individual								
Moisture	1.8108	0.5732	32	<0.001	0.5732			
Sand	1.8699	0.6116	32	<0.001	0.6116			
Salinity	1.9340	0.3143	16	<0.001	0.3140			
Elevation	1.9296	0.6826	35	<0.001	0.6825			
Reduced	1.7786	0.8989	51		0.7369	0.1026	0.0410	0.0184
Partial								
Moist’/SSE	1.7786	0.0540	3	<0.001				
Sand/MSEa	1.7786	0.0475	2.7	<0.001				
Salinity/MSSb	1.7786	0.0700	3.9	<0.001				
Elev’n/MSS	1.7786	0.1143	6.4	<0.001				

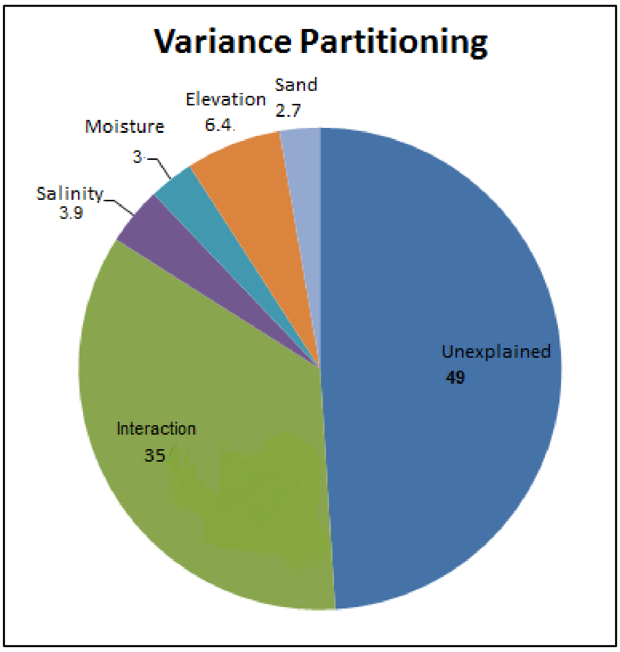


Figure 5-7 The portion of variance in the foraminifera data set that is unexplained, explained by the unique proportion for elevation, moisture content, salinity, sand content and the interaction between them ($p = < 0.001$).

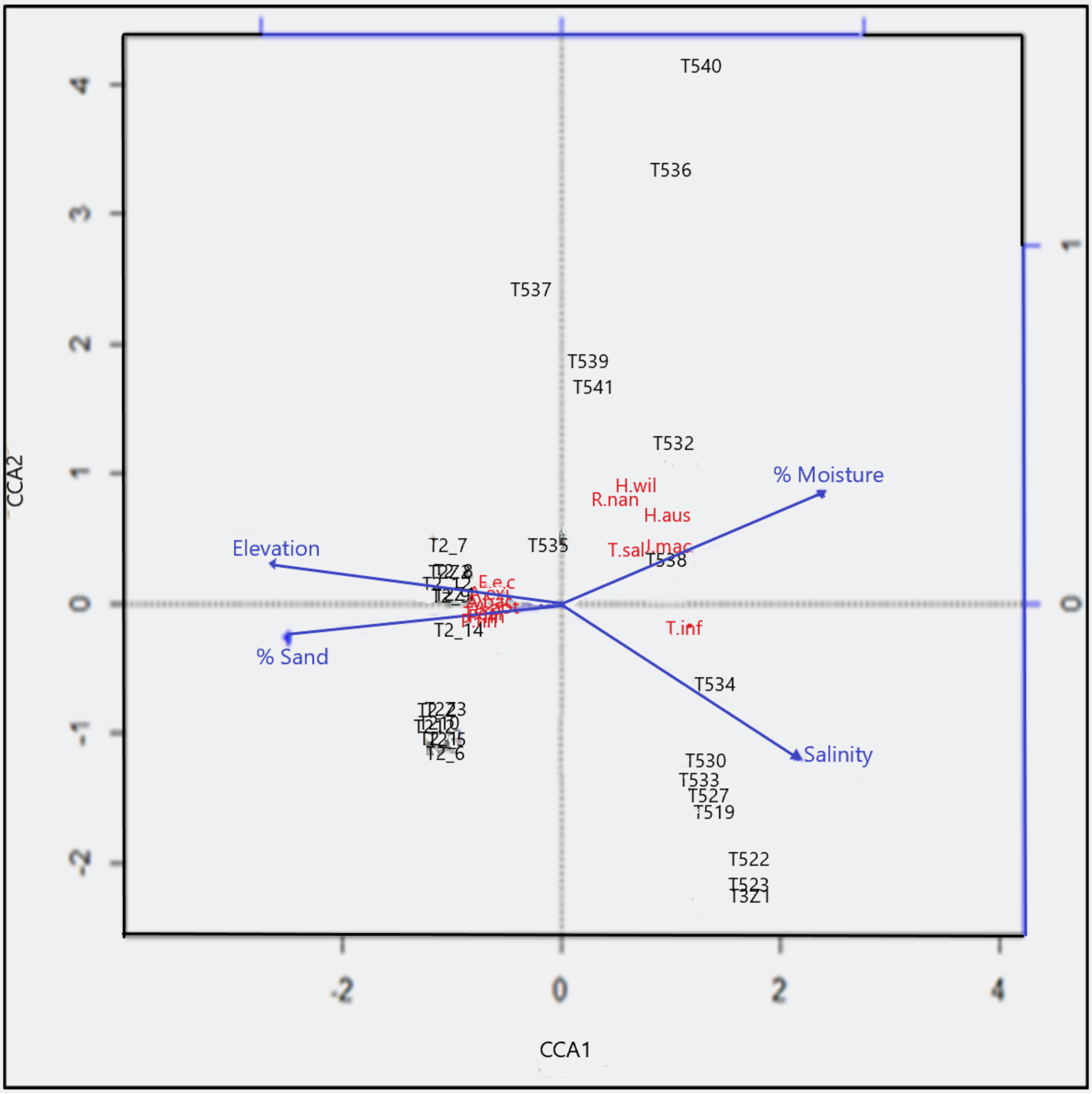


Figure 5-8 CCA biplot of species and samples for the reduced environmental variables model.

The direct gradient analysis retains the patterns of species and sample variance of the NMDS ordination, with tidal flat samples to the right, and salt marsh samples to the left on axis 1, although the distance between groups is greater (see Figure 5-8). This is expected because CCA maximises the variance between samples (ter Braak and Verdonshcot, 1995). The separation of tidal flat samples and salt marsh samples are explained by axis 1, on which inundation period and % sand content trend in similar directions. Variation on axis 2 relates to sediment moisture and salinity.

5.3.3 Transfer Function

Model Performance

Four outliers that were removed from the training set included three from the high point of the measured elevation (42, 43, 44) gradient that had low individual counts of dead *Jadammina macrescens* and *Trochammina inflata*. One outlier was a sample from the mid marsh (28) containing a high relative abundance of *Milliammina fusca* with *Quinqueflora seminula,* and *Ammonia aoteana* and was considered allochthonous due to its proximity to a small, infilled tidal creek where *Samolus repens* was dominant. The dominance of *S. repens* indicates that this location is wetted daily (see 3.4).

For the remaining 41 samples, weighted averages partial least squares regression provided the best performing model for the full intertidal, the seaward edge to low marsh, and tidal flat only data sets. Component 1 was shown to produce the most realistic reconstruction compared to component 2 which provided a slightly higher R^2_{boot} and smaller RMSEP (see Table 5-3). However, the reconstruction from component 2 placed several core samples that contained low intertidal molluscs above MSL, and was rejected on that improbability

Table 5-3 WA – PLS regression parameters for the 3 intertidal range transfer functions in this study and the salt marsh FBTF developed by Callard et al., (2011). (WA-PLS C1 – weighted averages partial least squares component 1, $RMSEP_{boot}$ – root mean squared error of prediction bootstrapped.

Model and elevation range	WA-PLS C1	Bootstrapped r^2	$RMSEP_{(m)}$
Full intertidal -0.46 to + 0.58 m	1	0.76	0.16
41 samples, 12 species and 1 genera	2	0.81	0.15
Salt marsh 0.13 to 0.58 m	1	$0.42_{(jack)}$	0.13
43 samples (Callard et. al 2011)	2	$0.65_{(jack)}$	0.10
Tidal flat 0.11 to – 0.46 m	1	0.27	0.16
16 samples		0.28	0.20
Tidal flat / low marsh -0.46 to + 0.34 m	1		
33 samples	2	0.70	0.16
		0.74	0.16

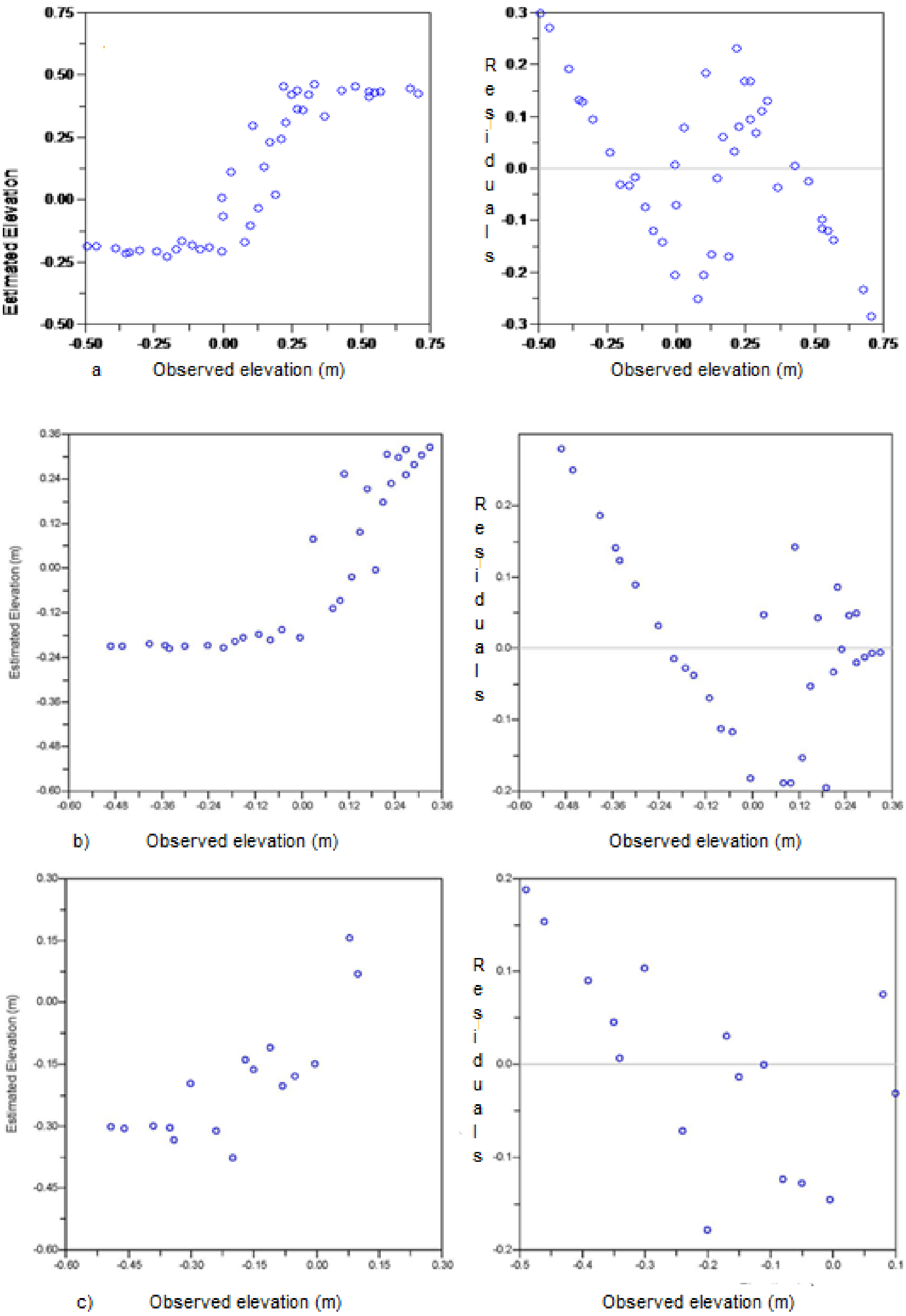


Figure 5-9 Observed elevations against inferred elevation for the 3 elevation gradient sections, by WA PLS regression component 1 on the left, and the residual errors on the right, **a)** full intertidal gradient, **b)** tidal flat to low marsh and, **c)** tidal flat only, showing that only segments of each section are linear relationships.

Table 5-5 Species elevation optima and tolerances for the reduced training set.

Species	Elevation optima (m)	Elevation tolerance (m)
Trochammina inflata	0.46	0.22
Jadammina macrescens	0.36	0.12
Haplophragmoide wilberti	0.25	0.16
Haplophragmoides australensis	0.38	0.17
Milliammina. fusca	-0.22	0.38
Excavatum excavatum	-0.17	0.27
Quinqueloculina seminula	-0.13	0.16
Ammobaculites agglutinans	-0.15	0.31
Ammobaculites exigus	-0.10	0.12
Ammonia spp.	-0.12	0.19
Reophax nana	-0.11	0.04
Protoschista findens	-0.14	0.20

In the reduced training set *R. nana* is most constrained to elevation, with an elevation optimum of - 0.11 m, and a tolerance range of 0.04 m. The optima for other tidal flat species occur around the upper tidal flat section, with tolerances ranging from 0.36 m for *M. fusca*, to 0.12 m for *A. exigus*. *M. fusca* has the largest elevation tolerance of 0. 36 m. In vegetated sections *J. macrescens* had the narrowest elevation tolerance of 0.12 m and *T. inflata* had the highest with 0.22 m.

5.3.4 Core Calibration

Core 2bT5LB was recovered from the mid marsh along transect 5 at Luttrells Bay at an elevation 0.41 cm relative to MSL. Fossil recovery was sporadic and samples did not exceed 40 individuals. The full intertidal model FBTF was selected for core calibration from which *M. fusca* was excluded from the training set, firstly because it was found to span the entire tidal flat to the mid marsh. It yielded a large elevation tolerance of almost 25% the entire measured intertidal range. Secondly *M. fusca* has a finely cemented, large and relatively hollow teste which apparently does not preserve well. Two *Ammonia* species were identified in the modern analogue. *Ammonia beccari* and *Ammonia aoteana* were discerned by their size and dorsal and umbilical surface features (see appendix 1). However, the teste features of fossil *Ammonia* were not discernible, and so all individuals were grouped into the genus *Ammonia* in the transfer function. The grouping also resolves the problem of including two species that are regarded by some as conspecific (Hayward et al., 2004).

Six significant biostratigraphic zones were identified using a broken stick model of fossil foraminifera assemblage variance, (see Figure 5-10). The zones were consistent with the a priori facies divisions that are described in 5.3.5 and discussed as palaeoecological phases (see Figure 5.11) in Section 5.4.1.

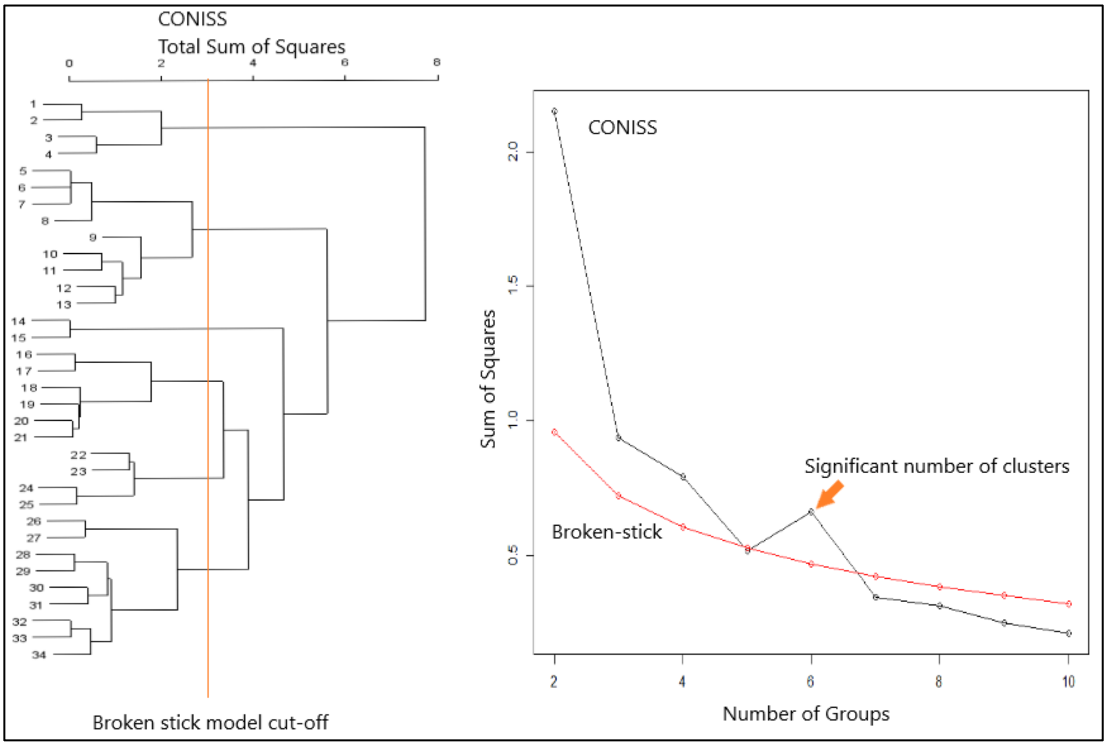


Figure 5.10 Broken stick model and CONISS cluster analysis of fossil foraminifera showing six significant clusters of species (the dendrogram is added to the reconstruction diagram Figure 5.12).

5.3.5 Facies Descriptions

Facies 1 162 – 140 cm

From the core base at 162 to 120 cm, a low intertidal mud flat facies is indicated by abundant, intact, macrofossils that occur commonly on the modern tidal flat below MSL in Luttrells Bay, and by some open marine intertidal species. The sediment is composed of a matrix of fine quartz sand, silt and shell hash with some minor lithic components. The core showed little variation in sediment colour and texture. Molluscs declined from 125 to 120 cm and were absent for the rest of the facies to110 cm. This hiatus in molluscs was observed in C1T5LB, however there was no salt marsh facies above it (see section 4.5) compared to the core examined here. Foraminifera recovery was low, and dominated by *A. agglutinans* that were smaller than those in the modern analogue. *Ammobaculites exigus*, and *Protoschista findens* did not occur in all samples but when present, occurred with relatively high abundance.

Facies 2 110 – 100 cm

This thin salt marsh facies is evident at 110 to 100 cm. It is characterised by an absence of molluscs and an absence of shell hash or lithic components. *Sarcocornia* fragments were present with abundant low salt marsh foraminifera fossils, dominated by *T. inflata* and *H. wilberti*. It appears in the reconstruction diagram (Fig 5-10b) with sharp upper and lower facies contacts.

4.3.3 Facies 3 100 – 55 cm

Overlying the late Holocene salt marsh facies is another low intertidal facies with fewer molluscs that the first intertidal facies. The molluscs are less intact, and appear more weathered than those in the mollusc facies below it, suggesting that some have been transported, rather than being in their life position. Fossil foraminifera occur with less than 20 individuals, dominated by *Ammobaculites agglutinans* with less frequent *Ammobaculites exigus* and *P. findens*. *Ammonia* spp. are minor components of the fossil assemblage in this facies. However, at 90 cm depth, micro fossils are essentially absent, except for infrequent, small *A. agglutinans*.

Facies 4 55 –47 cm

A salt marsh facies was visually evident by colour, texture and presence of *Sarcocornia* root which extended to from 55 cm to 47 cm depth in core but were not densely matted. It was devoid of molluscs and fossil foraminifera were rare, and only consisted of *A. agglutinans* and infrequent *T. inflata*. The facies was interpreted as representing the initiation of salt marsh. Its modern analogue is salt marsh seaward edges with sparse pioneer vegetation, rather than sparse and drowning vegetation such as occurs on eroding shorelines.

Facies 5 47 – 0 cm

The salt marsh initiation facies is overlain by a modern salt marsh facies comprised largely of dense *Sarcocornia* root mat in fine sandy silt. Foraminifera were common, and were dominated by *T. inflata* and *J. macrescens,* with occasional *H. wilberti.*

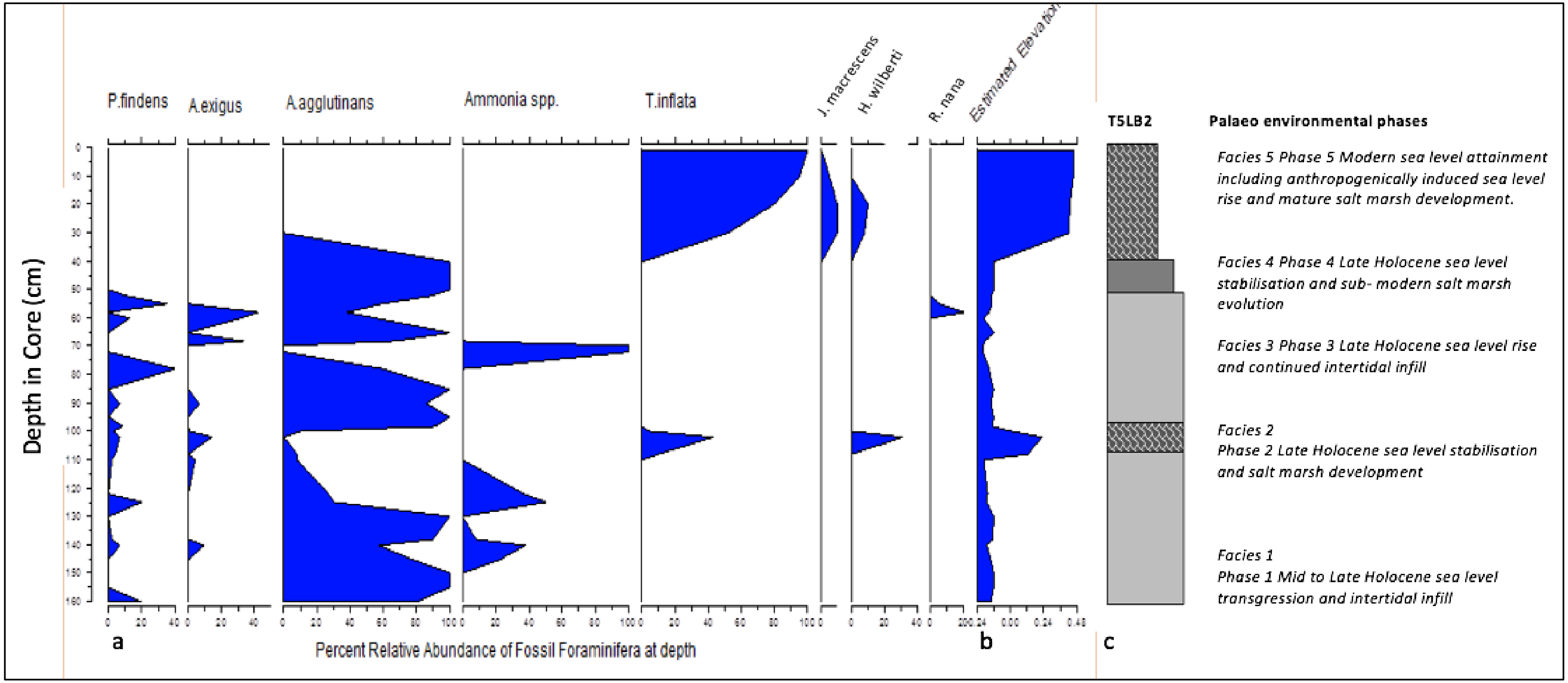


Figure 5-11) a) Weighted Averages – Partial Least Squares, Component 1 transfer function and, **b)** reconstruction of the estimated elevation of fossil samples in core T5LB2 and **c)** facies diagram.

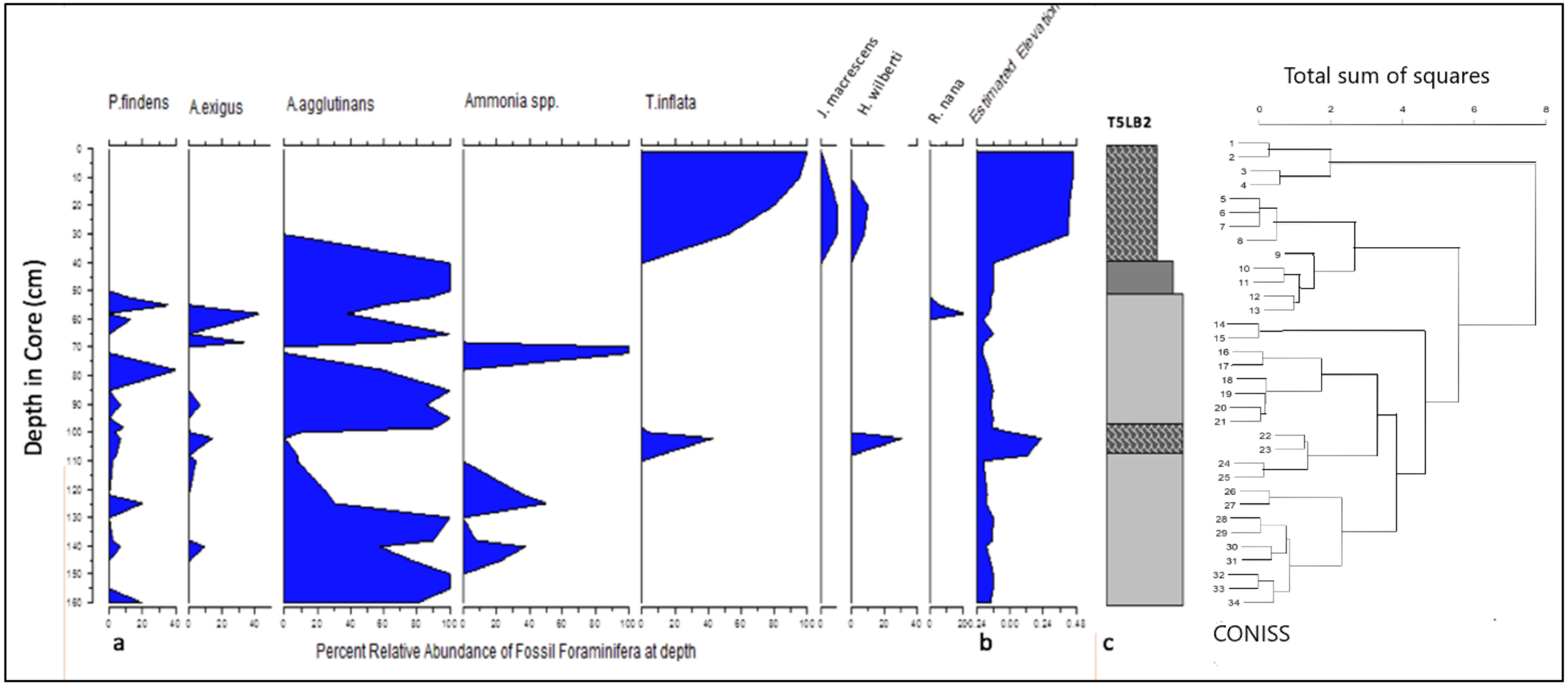


Figure 5.12 The transfer function and reconstruction with stratigraphically constrained cluster analysis showing the variance in fossil foraminifera assemblages, against the palaeo environmental phases interpreted from lithostratigraphic and macro and micro fossil evidence.

5.4 DISCUSSION

Elevation relative to mean sea level is the strongest variable influencing species distribution across the full intertidal gradient in Luttrells Bay. However, the inundation period controlled by this elevation co-varies with salinity, moisture content and sediment texture as in other salt marshes (Horton and Edwards, 2006). Tidal inundation period controls sediment salinity because NaCl concentrates differentially with evaporation during tidal exposure (Adam, 1990); elevation controls drainage potential of the sediment because sediment texture varies differentially with tidally transported and deposited allochthonous material (Fagherazzi et al., 2006; Fagherazzi and Wiberg 2009; Fagherazzi et al., 2012) and moisture content is similarly dependent on this relationship, and freshwater inputs.

The full intertidal FBTF RMSEP of ± 16 cm, shows a reasonable, yet relatively low level of accuracy in comparison to the salt marsh only FBTF of ± 10 cm (Callard et al., 2011). Although the model RMSEP is reasonable and the $r^2 = 0.76$ an acceptable coefficient of determination, the result cannot be accepted due to clear patterns in the WA-PLS residual plot, indicating a poorly performing model (see Figure 5-9a). Samples from around the highest tide mark and the lowest tide mark produced the largest residuals.

The tidal flat transfer function does not offer a level of accuracy suitable to the development of high precision SLIPs to reconstruct intertidal infill, due to low WA-PLS C1 R^2_{boot} value of 0.27. The poor strength of the relationship between elevation and species variance along the tidal flat indicates that caution should be applied to a reconstruction using the full intertidal model, because its strength is clearly lent by the precision provided by the low to mid marsh sections, predictability being absent outside this range. The trends are similar to the sectioned macro tidal mangrove elevation FBTFs; in that the strength of the full model is lent by species turn over in the low to mid vegetated sections (Woodroffe, 2006). In this case the micro tidal full model RMSEP should be interpreted as smaller, and R^2_{boot} higher, than the macro tidal FBTF, because the elevation ranges of all zones is smaller. Notwithstanding the limitations of the full intertidal FBTF, it broadly indicates tidal planes that fill the gap between 3700 yr BP and 650 yr BP.

5.4.1 The Reconstruction

Facies descriptions for sediment texture and molluscs correlated well with the WAPLS Component 1 FBTF reconstruction shown in Figure 5-11a. Most evident are two sharp changes from low intertidal species, to mid to higher intertidal species. The stratigraphic diagram (Figure 5-11b), shows them as sharp facies contacts which could be interpreted as a record of rapid change in elevation relative to MSL, or a significant erosion or deposition event (Boggs, 1987). However, it is more likely that the contacts reflect the sharp changes in assemblages across the tidal flat to low salt marsh. This is supported by the distance between tidal flat and salt marsh assemblages in ordination space, compared to the more transitional change in species across the salt marsh groups, see Figure 5-5. The changes may be exaggerated because samples from 50 – 90 cm depth in core, were either barren, or had low counts of *Ammobaculites agglutinans,* and only one modern analogue sample at the seaward edge of T2LB contained 100% *A. agglutinans.* However, in reality the transition from tidal flat to marsh seaward edges is geomorphologically abrupt, having been shown to occur at a mean elevation of 14 ± 3 cm at Luttrells Bay, and variably around MHW across different tidal, climatic and geomorphic settings in Tasmania (see Chapter 3.4.1).

The base of the modern salt marsh facies occurs at 0.55 m depth in the core, or 0.15 m below PMSL, indicating that at the time of its deposition, MSL was at around 0.29 ± 0.03 m below present level. This is consistent with the 650 yr BP SLIP at the base of salt marsh facies in Watch House Bay in Little Swanport at – 0.28 m (Gehrels et al., 2012). Based on the strong correlation between facies elevations between bays, facies below the modern salt marsh in Core 2bT5LB are assigned an age older than 650 yr BP. The base of the 162 cm core is at 121 cm below PMSL with an inferred age of approximately 4000 yr BP, based on its depth proximity to SLIPs reported previously in Luttrells Bay (see 4.3.2). Relatively constant and slow rates of sea level rise during the period are assumed to develop an inferred chronology for palaeo-elevations of MSL plotted on Figure 5-13. Palaeo-elevations are inferred for each facies from its sedimentological features and macro fossils, and are correlated with the FBFT reconstruction (Figure 5-9b). The inferred geomorphic processes are described as geochronological phases of estuarine evolution after Sloss et al. (2006), and include additional detail on salt marsh evolution.

Phase 1 Late Holocene estuarine infill and slow sea level rise from approximately 4000 to 2400 yr BP. Fossil foraminifera, molluscs and, sediment texture of facies 1, indicate marine transport of sediment landwards, with minor terrestrial input. The high density of intertidal molluscs suggest sediment transport is constant, with a rate of sea level rise slow enough for in-situ, low intertidal mollusc communities to thrive and dominate the macro fossil assemblage. These intertidal marine molluscs were present at greater depths in the same low intertidal facies recovered previously from Luttrells Bay (see 4.3.2). The absence of molluscs at the top of the facies indicates that elevation was approaching slightly above MSL, which in Luttrells Bay is an elevation zone existing in a state of dynamic instability because, if it accretes above tidal flat elevations, salt mash will develop (see 3.5.1).

Phase 2 brief sea level stabilisation or still stand from approximately 2400 – 2200 yr BP. Salt marsh fossil foraminifera, sediment texture and the absence of molluscs suggests a period of sea level stabilisation after the surface elevation of the intertidal flats reached around 14 ± 3cm above palaeo MSL (see 3.4.2). The 10 cm facies suggests that salt marsh establishment occurred beyond an initiation phase which in Tasmanian modern environments can be brief and not lead to established mature marsh (author’s observation), so relatively stable sea levels are likely to have occurred. The palaeo salt marsh facies occurs midway between the core base (approximately 4000 yr BP), and the base of modern salt marsh (approximately 650 yr BP). Based on the assumption that rates of sea level rise have been slow and constant the palaeo salt marsh facies was deposited around 2400 years BP, and so MSL was around 40 – 50 cm below present levels at this time.

Phase 3 recommencement of slow SLR and continued estuarine infill. Low intertidal foraminifera, sediment texture and a mixture of broken and intact marine and intertidal molluscs indicates sediment transport landward, with a recommencement of sea level rise from around 2200 yr BP, to around 650 years BP. The declining mollusc numbers toward the top of the facies indicate that the elevation of deposition is again reaching just above palaeo MSL.

Phase 4 salt marsh establishment and Phase 5, modern, mature salt marsh evolution. Sediment texture, *Sarcocornia* roots and an absence of molluscs indicates the initiation of salt marsh at around 650 yr BP, which continued to develop through to mature marsh in Phase 5, indicated by typical salt marsh foraminifera fossils.

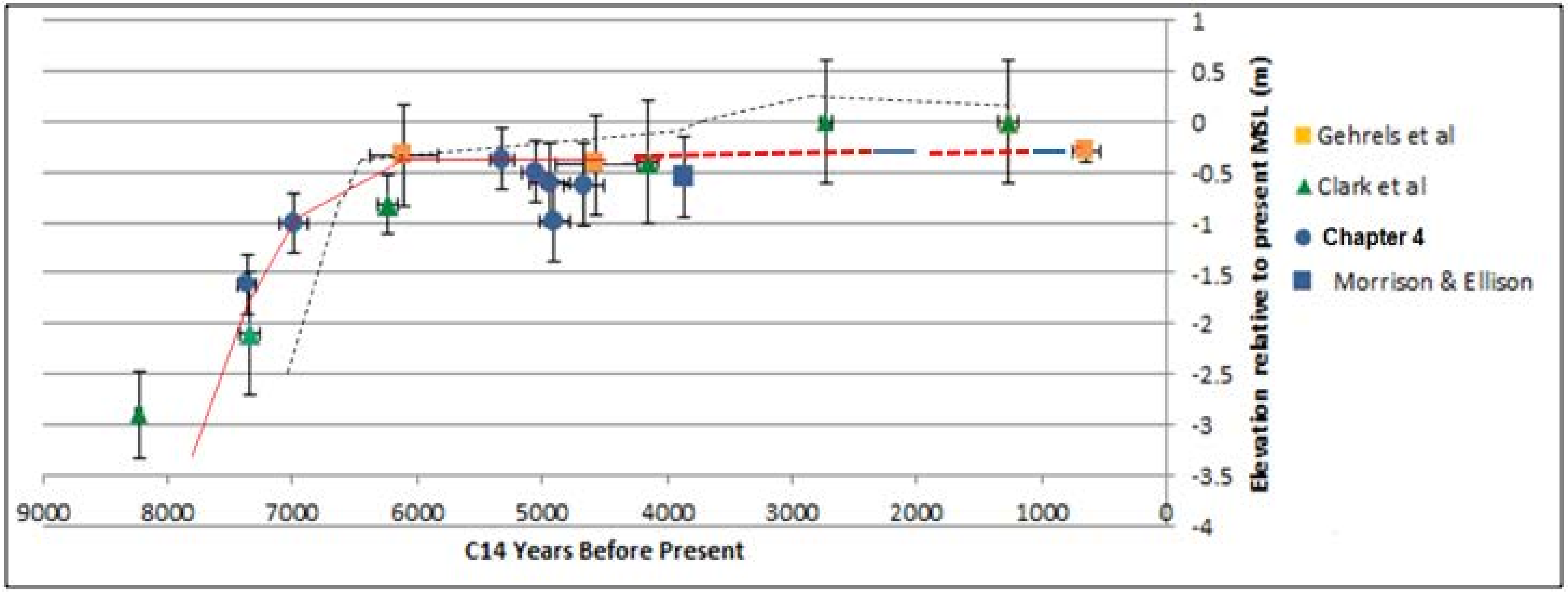


Figure 5-13 Holocene sea level curve for south eastern Tasmania including the likely nature of sea level change from around 4000 yr BP to 650 yr BP inferred from this study. The red dashed line is the inferred rate of sea level rise that was very slow, including a period of sea level stillstand (blue line), following the same trend as the mid Holocene from 6000 to 4000 yr BP. At around 2000 yr BP, sea level rise accelerated slightly to around 650 yr BP, when the present salt marsh began to develop. This period requires confirmation with precise sea level indicators which remain as a gap in the sea level record.

5.5 Conclusion

Despite the full intertidal FBTF’s limitations, the use of additional ordination techniques and explicit treatment of the surface training set, as recommended by Wright et al. (2011), extended its capacity to a broadly indicative model, suitable for the location and research question. Furthermore, sampling of very low marsh and seaward edge geomorphological variation did provide additional modern analogues. Considering the indicative meanings provided by salt marsh facies, the macrofossils located at the top of the low intertidal facies should represent MSL position well and, thus, be useful for ^{14}C dating to provide geochronological constraint.

The reconstruction elucidates the nature of sea level change during the late Holocene to around 650 yr BP, in particular excluding any excursion of sea level above present levels which had been previously suggested (Murray-Wallace and Goede, 1991; Haworth et al., 2002; Clark et al., 2011; Gehrels et al., 2012). The reconstruction showed that from 4000 yr BP levels rose from around 0.5 to around 0.25 m below present levels at 650 yr BP, equating to a very slow rate of sea level rise of around 0.08 mm per year, assuming constant rates. The trend also departs from the glacio - hydro isostatic adjustment model predictions described by Gehrels et al. (2012). The indications presented in this study support the SLIPs presented in Chapter 4 and provide further input for new glacio-hydro isostatic modelling for the Australian region.

Chapter 6

Late Pleistocene and Holocene intertidal stratigraphy of Boullanger Bay, north western Tasmania: Seeking an early to mid Holocene sea level record

6.1 Introduction

Precise Holocene sea level reconstructions in Tasmanian micro tidal estuarine locations in south eastern Tasmania have been limited to the late Holocene. An intertidal stratigraphic search around Tasmania revealed a lack of deep salt marsh deposits, as exemplified at Luttrells bay and other south-eastern Tasmanian estuarine sites (Clark et al., 2011; Gehrels et al., 2012; Morrison and Ellison, 2017), see Figure 6.1. The shallow depth of salt marsh deposits is a likely consequence of micro tidal regimes, slowly rising sea levels during the mid to late Holocene (see Section 4.5 and 5.5), low organic sediment production (Beasy and Ellison, 2013), low terrigenous sediment supply from low rainfall, low relief, ancient geologies (Twidale and Campbell, 1988), and negligible recent marine sediment transport to the coastline (Passlow, 2008). Sediment transport began to diminish in the later stages of the Last Postglacial Marine Transgression (Nichol et al., 1994).

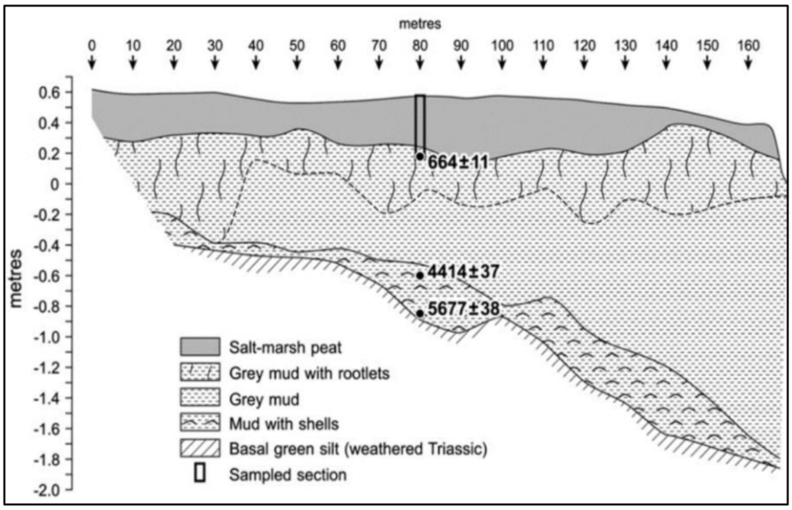


Figure 6.1 Stratigraphic profile typical of intertidal saltmarsh at Luttrells Bay Tasmania (adapted from Gehrels et al., 2012), showing a shallow and young salt marsh deposit that overlies intertidal muds and a muddy transgressive sand sheet typified by abundant shallow intertidal molluscs. The marsh is less than 664±11 years old (calibrated years before present).

The present study examines salt marsh and related intertidal deposits in north western Tasmania. The site has the greatest tidal range in Tasmania, and therefore the greatest potential to hold longer records than the previously studied micro tidal site (Chapters 4 and 5), as is the case in meso to macro tidal areas in Europe (Flemming and Bartholoma, 1995; Ruiz et al., 1998; Castillo et al., 2000; Bartholody, 2012), and the USA (Hill and Shearin, 1970; Gehrels 2002; Kemp et al., 2015). In the study region, the estuaries and bays are well-filled with sediment derived from the shelf, during and after the post glacial marine transgression (Davies and Hudson, 1987). Since sea level rise rates slowed after 6000 yr BP, the erosion of coastal bedrock and accession of biogenic carbon have been the primary sediment sources, with little contribution from rivers (Davies and Hudson, 1987). A tide-dominated sediment mobility regime currently exists (Passlow et al., 2008). With sufficient sediment supply the sea level signal should be captured in the sedimentary record (Morrison and Ellison, 2017).

Boullanger Bay is a large bay that lies on the lee side of Tasmania’s north western tip. Robbins Island lies to the north of a passage between it and the mainland. The coastal hinterland is extensive due to its position within the shallow and broad Smithton Basin. This basin formed during the Late Precambrian and is variably overlain with Quaternary sediment (Calver et al., 1995). The shallow basement and high rainfall produce a low-lying coastal flood plain, dominated by wet coastal scrub vegetation communities (Seymour and Calver, 1995). Previous palaeoecological studies in the Smithton Basin, around 50 km south west of Boullanger Bay document freshwater marl and swamp peat deposits that developed under the influence of fluctuating groundwater conditions in palaeo-floodplain and sand dune ridge and swale environments over the last 50 000 to 11 000 years (Gill and Banks, 1956; Colhoun et al., 1982).

Despite its extent, and geomorphic importance, Boullanger Bay’s intertidal landforms have not been studied. Field reconnaissance revealed an extensive black, silty fine sand deposit underlying the modern eroding saltmarsh (Figure 6-2a and 6-3a). Due to its stratigraphic position, it was inferred to be an older, deeper, salt marsh facies. Between the two inferred salt marsh facies was another, composed of fine sands inferred to be a transgressive sand sheet, similar to that underlying the south-eastern salt marshes (see Section 4.4 and 5.4).

The aim of the present study is to determine the age of the deeper black silty fine sand (BSFS) deposit, and the spatial organization of the stratigraphic units at Boullanger Bay. The objective is to understand the geomorphic evolution, and potential to provide a precise early to mid Holocene sea level record, of this deposit.

6.2 Methods

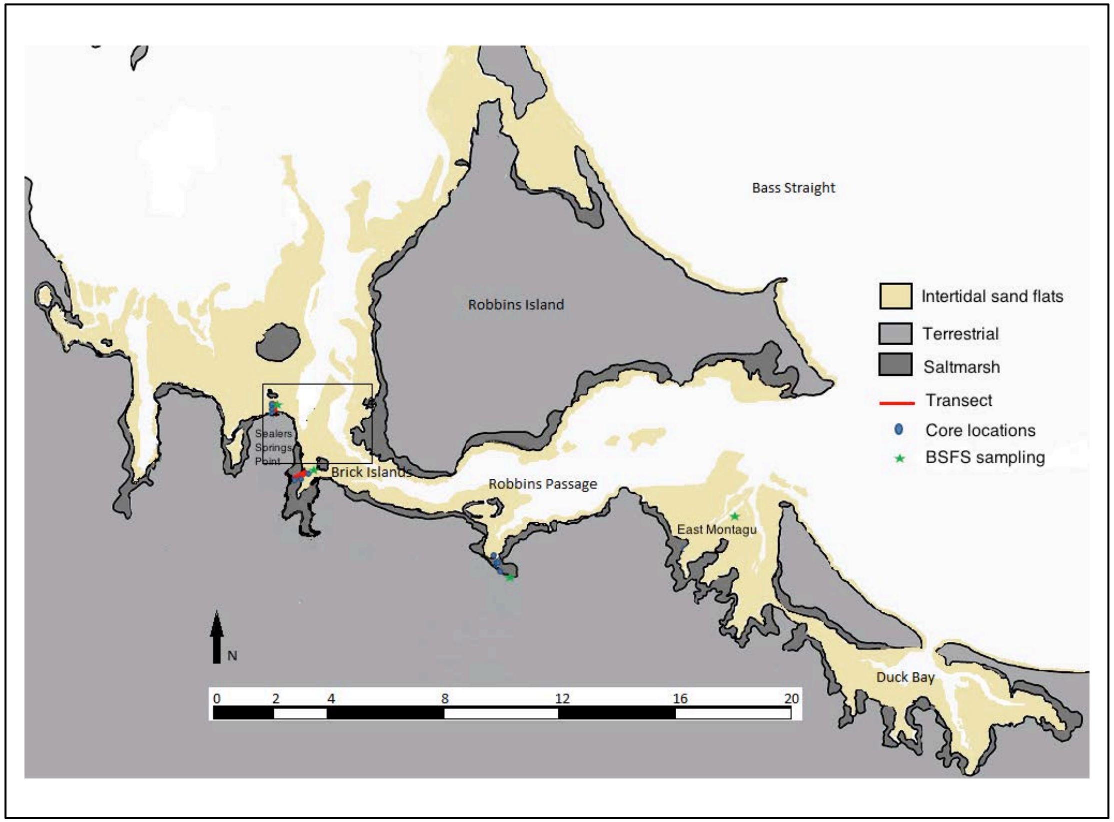
6.2.1 Study site

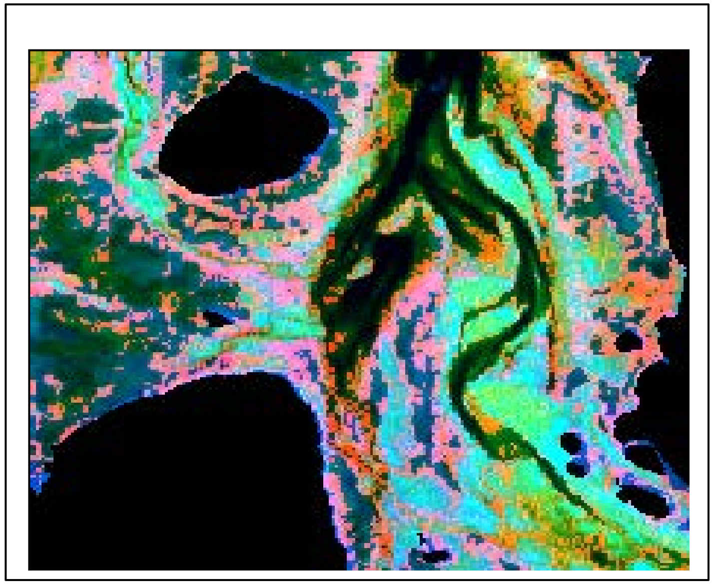
Salt marshes in Boullanger Bay are the only open coast type in Tasmania, owing to the quiescent wave conditions created by intertidal sand flats, that, in parts, are over 2 km between low tide mark and the fringing salt marsh. Three mature salt marsh sites were selected in different wave energy zones; open coast, semi sheltered tidal passage inlet and sheltered passage.

Site 1: Sealers Spring Point on the open coast is the most exposed site (see Figure 6.2) with actively eroding marsh seaward edges (see Figure 6-3a). Micro-cliffing was extensive along a seaward edge dominated by *Juncus kraussi* and *Distichlis distichophylla*, with subdominant *Sarcocornia blackiana*. This community is typically mid-marsh (see 3.3.2). The terrestrial edge was dominated by *Melaleuca ericifolia* swamp forest. The site was selected because the profile of the BSFS deposit could be observed from an accessible, intertidal to shallow subtidal channel and across the upper intertidal flat (see Figure 6-3a, b and c), enabling the stratigraphic relationship between facies to be recorded.

Site 2: Brick Islands lies within the western inlet of Robbins Passage, so has some protection from open coast energy (see Figure 6.4). The presence of five distinct vegetation types distributed across low, mid and high marsh (see Section 3.3.2) and one major tidal creek, indicated that the marsh was mature (Davis and Fitzgerald, 2004). Both accreting and eroding salt marsh seaward edges were evident (see Figure 6-6b). The BSFS deposit outcropped in the salt marsh tidal creek, extended out onto the intertidal sand flat creeks and was evident at eroding marsh edges (see Figure 6-6a and c).

Site 3, in Robbins Island Passage, lies at the confluence of a small freshwater and intertidal creek. Sheltered within the passage, it was most like most other intertidal sites in Tasmania (see Chapter 3, 4 and 5). Prograding low marsh was the dominant edge type, reflecting its sheltered setting. There was a seaward edge of *Sarcocornia quinqueflora* and *S. blackiana*.





6-2b

Figure 6-1a Boullanger Bay in North West Tasmania showing the extent of the intertidal sand flats and the fringing salt marsh and place names mentioned in text. Figure 6-2b is a highly processed multi spectral satellite image of the Sealers Spring Point area where the black silty fine sand deposit is exposed in the upper intertidal zone between Robbins Island and Sealers Spring Point. It shows the extent of the BSFS in pink. The mid-blue and blue/green areas are intertidal seagrass beds which overly the BSFS. The very dark green, pale green and orange indicates tidal channels, and black is terrestrial.

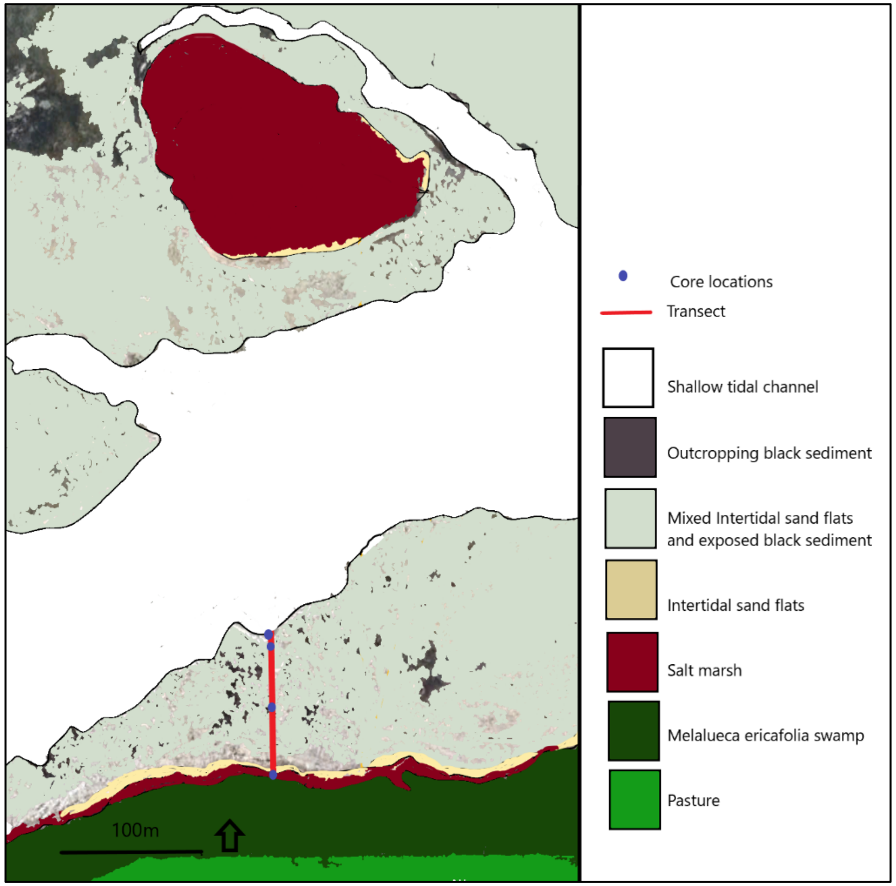


Figure 6.3 Sealers Spring Point site showing Transect 1 (T1SSP) location

a b c

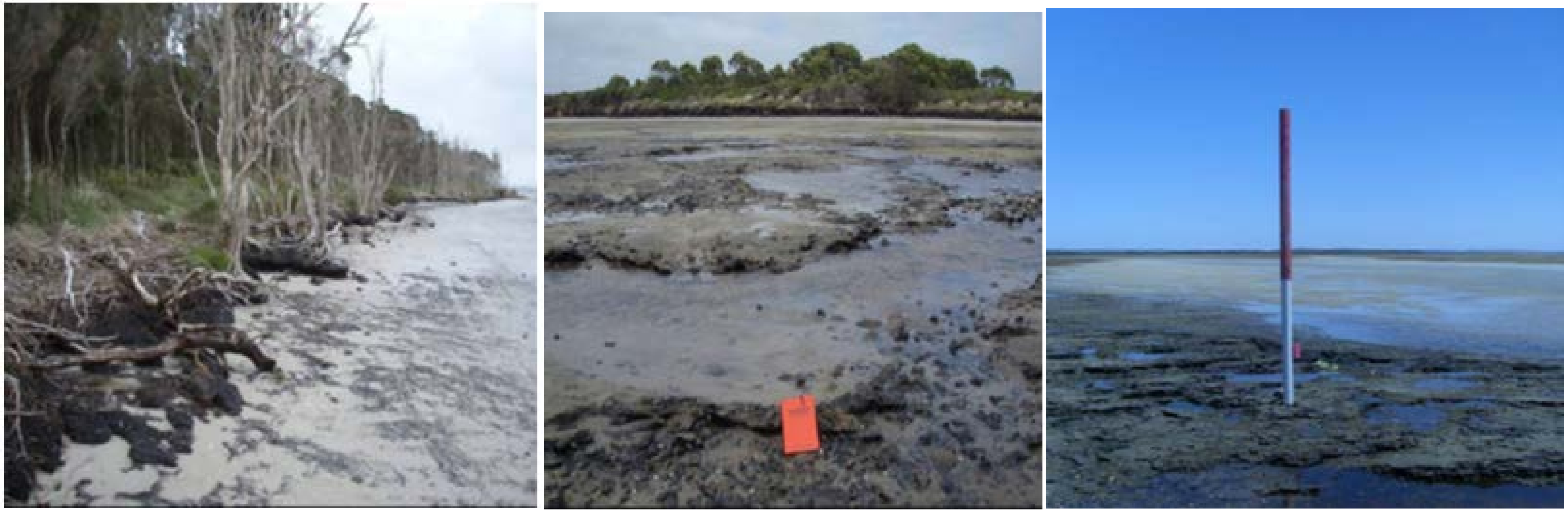


Figure 6-2 Transect 1 at Sealers Spring Point (T1SSP): **a)** the landward edge showing the BSFS underlying the modern *Melalueca* swamp forest and terrestrial edge of retreating salt marsh, **b)** the seaward edge of T1SSP showing the extent of the BSFS deposit and the fine sands overlying it, and **c)** the location of core 2 (T1SSP2) and the low to subtidal channel seaward of T1SSP.

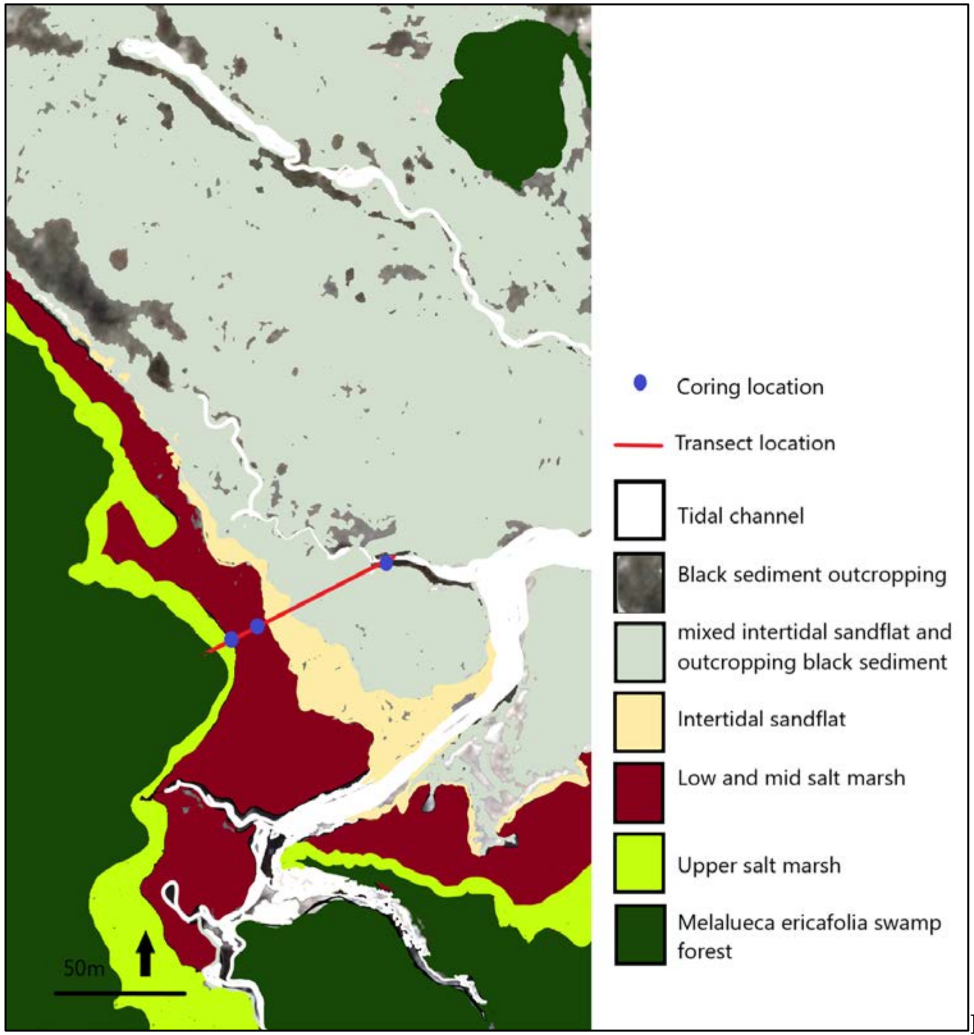


Figure 6.5 Brick Island site show transect 2 (T2BI) and core locations

a b c

Figure 6-3 a) The BSFS underlying the modern salt marsh and outcrops at the tidal creek at Brick Islands. **b)** T2BI showing prograding marsh edge and, **c)** the BSFS deposit outcropping in the tidal creek that dissects the sandflats is where T2BI terminates. Note in **c)** that faces seaward, the ocean at low tide is not visible on the horizon, illustrating the vast extent of the intertidal sand flats.

6.2.2 Transect surveys and sample elevation control

Precise locations and heights were obtained for core locations and surface stratigraphic units from transect surveys and spot heights tied to elevation relative to mean sea level. Elevation control points were created at Sealers Springs Point, using a Leica 1200 GPS and Leica TC407 Total Station differential GPS base station, deployed on the marsh terrestrial transition zone. Satellite data were downloaded for one hour and height resolution derived from the AUSPOS, and tied to the Australian Height Datum 1983 (AHD83). At Sealers Spring Point (SSP), one transect was deployed perpendicular to the main tidal channel from the terrestrial edge, across the marsh, and out onto the tidal flat, terminating in the tidal channel (T1SSP).

One survey control point was installed on Brick Island with an unobstructed view to the Brick Islands (BI) site on the mainland, and into Robbins Island Passage to the east, where spot heights of the BSFS, and the Robbins Island Passage (RIP) cores were recorded. On the mainland shore, a transect was deployed across the tidal flats to intertidal creek (T2BI), where the BSFS outcropped. The transect was surveyed for micro-relief, using a LeicaTC407 Total Station and crystal prism stave, and tied to the Brick Island control point by a closed loop traverse.

6.2.3 Coring

Coring with hand-operated devices failed to penetrate beyond the modern soft sediment salt marsh. To reach the deeper strata, a double-tube coring device, adapted from Tratt and Burne (1980), was built and used with the aid of a pneumatic hammer to force the outer tube into the sediment, so that compaction of the sediment caught in the inner tube was minimized. Coring was continued until refusal at each location, and the depth of the double tube device recorded and compared with the depth of the sediment core, to determine compaction by coring. Cores were sawn on location and split lengthways, logged for sediment color, texture and inclusions which collectively determined individual facies, and the BSFS was sub-sampled for carbon dating. The longest or most detailed core from each site was wrapped securely and transported to the laboratory for further analysis.

Cores were taken at T1SSP from the marsh terrestrial edge, the middle and lower tidal flats, and a shovel core at the main tidal channel. Seven shallow shovel pits were also dug sequentially along the tidal flat to identify the depth and extent of the inferred modern salt marsh facies underlying the tidal sand flats. Two cores were taken from Brick Islands on the salt marsh, and the depth of facies tied to those outcropping in the intertidal creek. Three cores were taken at Robbins Island Passage.

6.2.4 Stratigraphy

Cores were subsampled for macro and micro fossils, minerology, grain size and shape analysis for each facies. For T1SSPC4, the longest core, a total of 16 facies were observed and were categorized into 5 stratigraphic units based upon lithological and biological features evident from core logging in the field. A description of the prevailing climate during the formation of each unit was obtained from the literature (Colhoun et al., 1982; Mackintosh et al., 2006; Colhoun and Schimeld, 2012; Petherick et al., 2013). The BI and RIP stratigraphy was correlated with the SSP cores, using the same stratigraphic unit categories. A sample of the BSFS was taken from a major outcropping at East Montague to compare its lithology with the other three study sites.

6.2.5 Sediment Analysis

Two sets of sediment analyses were undertaken. Standard sieve analysis was conducted for grain size distribution parameters from samples in each stratigraphic unit. Mean and median grain size, degree of scatter, degree of symmetry and kurtosis were evaluated using inclusive graphical methods, based on the cumulative mass retained in sieve sizes -3, -2, -1, 0, 2, 3, 3.99 and < 3.99 phi, after Folk and Ward (1957). The results are expressed in phi units and were converted to microns (*u*m) for descriptions of facies lithology in Section 6.3.3. Sediment particle size classes, and % passing cumulative grain size distribution classes were adopted from Wentworth (1922), to describe lithological variation in the stratigraphic diagrams (see Figures 6-8, 6-11 and 6-13). The standard sieve analysis methods are particularly accurate for rounded sediments over 100 *u*m but are less accurate for smaller particles that might be platey, rather than round (Mingard et al., 2009)

To accurately record the distribution of particles finer than 100 *u*m, granulometric analysis was undertaken using a Saturn DigiSizer 5200 V1.09, which measures particle size in suspension and expresses the results as % volume of grain sizes, measured in microns. Minerology, inclusions and particle roundness was observed by compound microscope, and roundness comparisons made after Powers (1953).

6.2.6 Microfossil analysis

Samples for foraminifera were collected and wet sieved through a 125 and 63 μm sieve and both were scanned where fossils were rare, otherwise, only the 125 *u*m was scanned as per Gehrels (2008).Species taxonomy and preferred habitat range followed Hayward et al. (1999), Yassini and Jones (1995) and the intertidal modern analogue from Little Swanport (Chapter 5.4).

In T1SSPC4, facies that were barren of macro biological or lithological marine indicators, pollen was investigated. Samples were taken at 10 cm intervals and were prepared in accordance with methods outlined in Ellison (2008). A pollen diagram was prepared to show the relative abundance of each taxon, as a percentage of the total pollen sum (Figure 6-7). Trees, shrubs, ferns and aquatics were recorded, and other palynomorphs such as fungal spores, dinoflagellates and chlorophyllaceae were excluded. The environmental range for the dominant genus or species in each stratrigraphic unit were described from Kirkpatrick (1991), and DPIPWE (2015).

6.2.7 Core chronology

To constrain the age of the BSFS sequence, one sample containing plant fragment was taken for AMS ^{14}C dating from the top of facies two in core T1SSPC4 and one containing plant fragment was retrieved from near the top of facies three in T1SSC1, (see Figure 6-12). Samples were wrapped in aluminum foil and sent to

Beta Analytic Laboratories in Florida, USA. Both samples had ample carbon for analysis, (see Table 6-2).

6.3 Results

Grain populations were dominated by fine quartzose sands in all samples below the modern salt marsh facies. It was difficult to visually differentiate between sub-angular and sub-rounded. According to Powers (1953), the term subangual-subrounded is the appropriate class in this instance, and is often the case when working with granular or smaller sized particles. This term was thus applied. Loss on ignition ranged from 0.7% to 12.8%, the highest being recorded for the BSFS, (see Table 6-1).

Table 6-1 Grain size statistics and parameters for dominant mineralogy, roundness, mean and median grain size, mean particle dimension, degree of scatter, symmetry, kurtosis and grain size distribution s by cumulative % passing using the Wentworth scale, and % organic content for samples from each stratigraphic unit at each site.

Depth in core (cm) & SU	Roundness	Mean particle size ϕ	Median particle size ϕ	Degree of Sorting	Degree of Symmetry	Kurtosis	Particle size category (Wentworth)	Grain size distribution	Organic Content %
Sealers Springs Point Transect 1 Core 4 (T1SSPC4)									
43/5	Subangual/ subrounded	2.53	2.5	Moderately sorted	Strongly skewed toward fine particles	Very leptokurtic	Fine sand	Silty fine sand	2
65/4	Subangual/ Subrounded	2.45	2.45	Moderately well sorted	Fine skewed	Very leptokurtic	Fine sand	Silty fine sand	0.7
75/3	Subangual/ Subrounded	2.66	2.6	Moderately well sorted	Fine skewed	Very leptokurtic	Fine sand	Fine to very fine sand	12.8
185/2	Subangual/ Subrounded	2.62	2.6	Well sorted	Coarse skewed	Platykurtic	Fine sand	Fine to very fine sand	4.9
225/1	Sub rounded/angual	2.19	2.4	Moderately well sorted	Strongly skewed toward coarse particles	Leptokurtic	Fine sand	Fine sand	6
230/1	Sub rounded/angual	2.04	2.3	Moderately Well sorted	Strongly skewed toward coarse particles	leptokurtic	Fine sand	Fine to medium sand	10.1
Brick Islands Transect 2 Core 1 (CIBIT2)									
7/5	Quartz:Sub rounded/angual. Silt: amorphous	8.2	10	Moderately sorted	Coarse skewed		Very fine silt	Fine sandy silt	10.8

35/4	Sub rounded/angual	2.65	2.44	Moderately well sorted	Strongly skewed toward fine particles	Very leptokurtic	Fine sand	Silty fine sand	3.8
50/4	Sub Rounded/angual	2.64	2.45	Moderately sorted	Fine skewed	Very leptokurtic	Fine sand	Silty fine sand	2.2
100/3	Sub rounded/angual			Moderately well sorted	Fine skewed	Very leptokurtic	Fine sand	Fine sand	1.9
Robbins Passage Core 1 (RPC1)									
7/5	Clay and Silt: Amorphous Quartz: Sub rounded/angual	6.25	6.0	Moderately well sorted	Coarse skewed		Silt	Fine sandy silt	6.1
60/4	Sub rounded/angual	3.34	2.92	Very well sorted	Strongly skewed toward fine particles	Very leptokurtic	Very fine sand	Fine to very fine sand	1.9
130/4	Sub rounded/angual	2.83	2.75	Moderately well sorted	Strongly skewed toward fine particles	Platykurtic	Medium sand	Medium to fine sand	2.3

Fossil recovery was generally low, except in Robbins Passage cores where foraminifera were abundant. Foraminifera were present above the BSFS facies and dominated by agglutinated species which are known to preserve better than calcareous species (Hayward et al., 1999; Gehrels, 2002). Consequently, species diversity was low, so foraminifera fossil assemblages are used as a guide to their palaeo tidal position, rather than for inferring precise elevations.

Pollen recovery was low and all samples were barren of molluscs. Pollen numbers were extremely low in the first analysis so sampling and preparation was conducted a second time, to try and improve pollen recovery, but did not increase pollen numbers substantially. However, pollen were present in all facies.

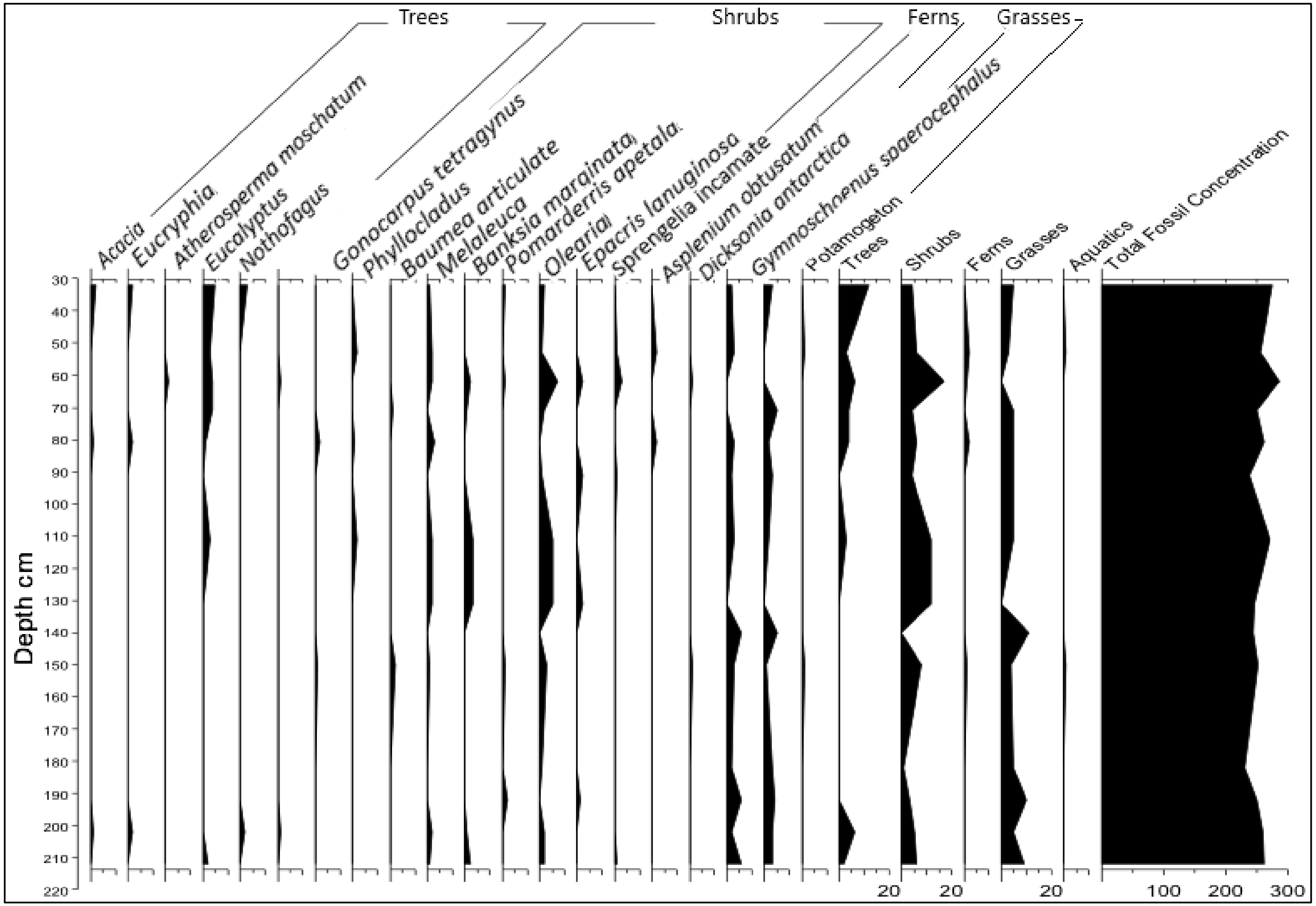


Figure 6-7 Pollen counts in T1C4SSP, below the modern salt marsh facies.

6.3.1 Core chronology

Two radio carbon dates constrain the age of the Facies 1 to 11 at Sealers Spring Point (Table 6-2). Dateable material from the facies overlying the fine black sediment deposit was sought, but due to contamination with both modern and older material, was abandoned. Living *M. ericifolia* and *Sarcocornia* root had penetrated the salt marsh, and microscopic analysis showed that particles of the underlying the BSFS were distributed throughout the overlying sediments. Agglutinated foraminifera testes had also incorporated the BSFS particles. The overlying sands were devoid of any macro-molluscs, or other suitable biological material for dating.

Table 6-2 AMS ^{14}C results constraining the age of the BSFS deposit.

Core and Sample code	**Material & pretreatment**	**Depth relative to present MSL**	**Conventional Radiocarbon Age**
T1SSPC1 SP1SWE	Plant material Acid/alkali/acid	-0.10 m	26,720 +/-180 BP
T1SSPC4 T1SPC3146-147	Peat Acid washes	-1.1 m	36,930+/- 400 BP

6.3.2 Core Stratigraphy

6.3.2.1 Sealers Spring Point (C4SSP)

Stratigraphic unit 1, facies 1, 230 – 207 cm

Facies 1 extended from 230 to 207 cm depth in the core and was designated a single stratigraphic unit, based on its color and the variation in sediment texture, compared to all other facies above it (see Figure 6-8). Mean grain size was 292 µm at 228 cm, being the largest of all samples in all cores. The grain population was moderately well sorted and strongly skewed toward coarse particles. The larger grains (> 250 µm), were mostly sub-angular. Colour and texture indicated derivation from dolomite which underlies the Quaternary sands (Calver et al., 1995).

The smaller grains were predominantly quartz (< 250 µm) that were sub-angular to sub-rounded and coated in black to brown silt. The matrix yielded an organic content at 10.1% which was obviously contributed by the silt component. Above the base of the core at 225 cm, the sediment remained moderately well sorted and strongly skewed toward coarse particles but the mean grain size decreased to 152 µm. Granulometric analysis showed a variable distribution of grain size by % frequency. The distribution at 220-221 cm was slightly bimodal, and at 217 and 210 was very well sorted very fine sands and was skewed toward finer particles (see Figure 6-8). The variation did not produce visible laminations. The base of facies 1 returned an aged of 36,930 +/-400 ^{14}C years BP.

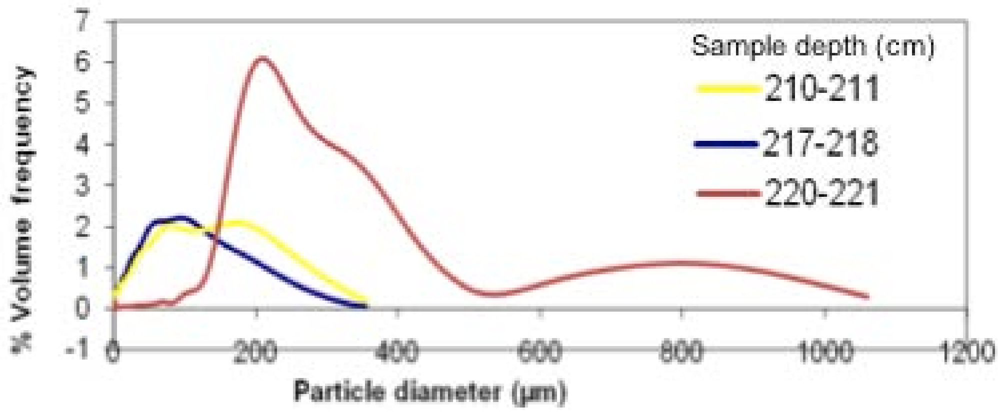


Figure 6-8 Particle size distribution for stratigraphic unit 1, facies 1 (220-221), and facies 2 of stratigraphic unit 2 (210-211 and 217-218) showing the difference in the sediment texture between the two facies.

Facies one was devoid of any marine fossils. Biological fragments of plants and seeds were common. Pollen at the base of the core was dominated by an unidentified species of the genus *Poa*, and *Gymnoschoenus sphaerocephalus* which occur adjacent but not in association in modern environments (Kirkpatrick, 1991). The latter is the dominant species of Buttongrass sedgeland communities in very wet, poorly drained, low nutrient quartzose soils in Tasmania (DPIPWE, 2015). However, *Poa* occurs on more fertile soils than *Gymnoschoenus sphaerocephalus* It can be succeeded by *G. sphaerocephalus* sedgeland in the absence of fire as a result of paludification (Kirkpatrick, 1991). Shrubs are represented by pollen dominated by *Olearia*, a genus that occurs from the coast to the alpine zone on well-drained ground (DPIPWE, 2015). *Melaleuca* pollen was present. This genus is a common dominant wetland shrub or tree on poorly-drained sites, occurring in association with *Eucalyptus* and *Acacia* in the swamps of the northwest (Kirkpatrick, 1991).

Stratigraphic unit 2 Facies 2 to 10 (207 – 96 cm)

A 7cm gradational contact occurred from facies 1 to facies 2, marking the beginning of stratigraphic unit 2 (SU2). The unit was distinguished on the basis of its black to brown color variation which produced definite laminations. The unit was separated into nine individual facies based on their variable depths and colour.

Lamination colour reflected the relative proportion of black silt, to fine sand. For example, facies 3 (185 – 180 cm) was composed of well sorted finely skewed, fine to very fine sands with a mean grain size of 160 *u*m and was very dark brown to black. The relatively high black silt content corresponded with a relatively high organic content of 6.4%. A sharp contact occurred to facies 4, (180 to 144 cm) that was comprised of distinct laminations of 2 to 4 mm depth. The mean grain size of the light brown laminations was 188 µm within a moderately well sorted grain population that was coarsely skewed with an organic content of 2.7%. The mean grain size of the darkest laminations was 150 µm. The particles were moderately well sorted and finely skewed. Organic content was relatively high at 9.7%.

The pattern of laminated and un-laminated facies repeated through facies 5 and 6 from 144 to 123 cm. Facies 7 was strongly laminated and the layers deepened to around 1.5 – 2 cm depth, from 123 – 113 cm. In facies 8, 9 and 10, the depth of the laminated facies decreased gradually within the consistently well sorted, fine to very fine sands. Granulometric analysis showed variation in the distribution tails, where some are coarse and some are fine (Figure 6-9). The variation reflects the alternating composition of the laminations over SU2. The dark laminations have a higher silt content, thus a finer tail. Notably the deepest sample from stratigraphic unit 2 at 200-201 cm was more coarsely skewed than any other sample. It reflected the gradational contact between SU1 and SU2, (see Figure 6-9).

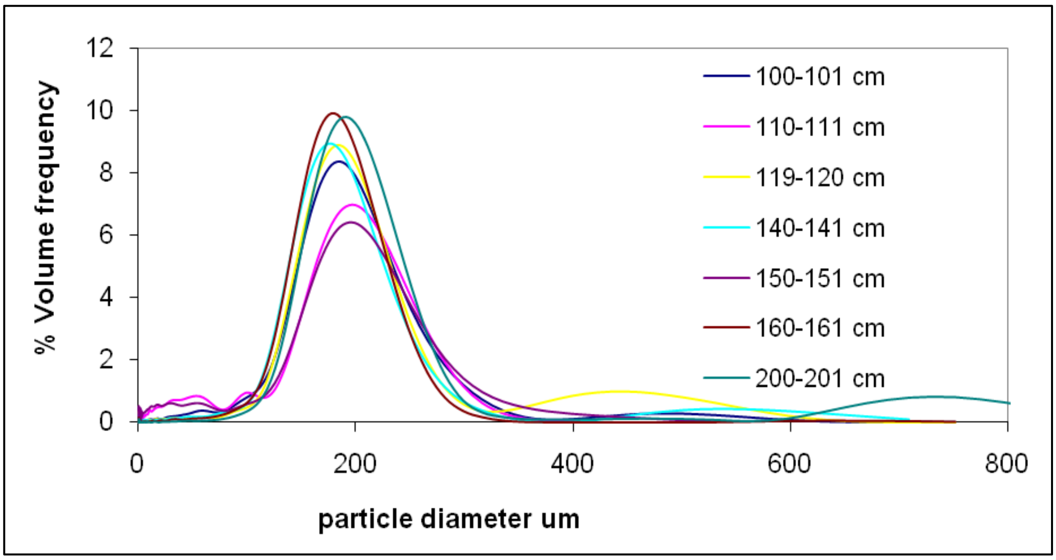


Figure 6-9Particle size distributions for samples from facies 2 through to 9 in SU2,

SU2 was barren of foraminifera or other marine fossils. From the base of SU2, the most discernable trend in pollen is the persistent dominance of *Poa* and *G. sphaerocephalus. Acacia*, *Eucryrphia* and *Nothofagus* pollen decline over the gradational contact between facies 1 and 2, and, except for *Eucalyptus*, none of the tree species occur again in SU2. Shrub pollen increased upwards through SU2, dominated by *Olearia and Epacris*. A low count of the freshwater aquatic genus *Potamogeton* (Kirkpatrick, 1991) was recovered from the middle of the SU2, but in very low numbers.

Stratigraphic unit 3, facies 11 (96 – 66.5 cm)

A 10 cm gradational contact to facies 11 was evident from a gradual decline in laminations, such that facies 11, was a monochromatic black sediment, designating it a stratigraphic unit 3 (SU3). Of all stratigraphic units under the surface saltmarsh, SU3 had the smallest mean grain size of 153 *u*m. It was a moderately well sorted, finely skewed fine to very fine sand. Size distribution by % volume shows the finer tail ranged from course to medium sized silt particles, (see Figure 6-10) and corresponds with the highest organic content of 12.8%.

The silt was observed microscopically as amorphous black to brown gelatinous particles when moist, which made up the sediment matrix, and coated the larger quartz grains. Laboratory handling revealed that when moist, the sediment matrix was firm, greasy and cohesive behaving as a semi-indurated mass. However, when air dry the mass disaggregated easily to a powdery form, and the fine black silt largely separated from the fine quartz sand particles.

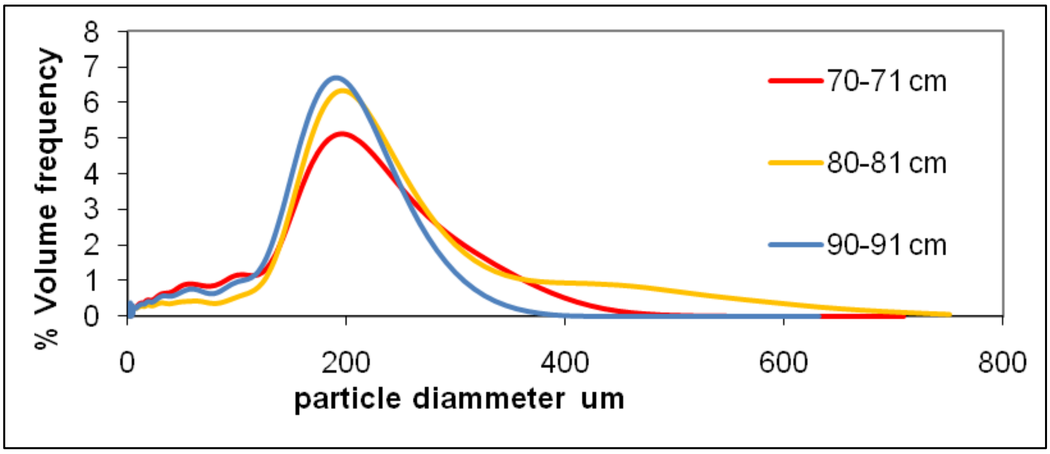


Figure 6-10 Granulometric Particle distribution by % volume for samples in stratigraphic unit 3, facies 11

The facies was barren of foraminifera or any macrofossils, indicating marine conditions. Grass pollen was dominant and highest in SU3 with *Poa* and *G. spaerocephalus*. *Melaleuca* was the dominant shrub. Wet forest *Acacia* and *Eucryphia* trees that were present in SU1 and absent in SU2, reappeared in SU3. The relative abundance of *Eucalyptus* trees increased in SU3 from SU2. *Asplenium obtusatum* and *Dicksonia antarctica* ferns were not present in SU2, but present in low numbers in SU3. While *A. obtusatum* is a coastal cliff species. *D. antarctica* is found widely in the understory of wet eucalypt forest and rainforest (Kirkpatrick, 1991).

Stratigraphic unit 4, facies 12-14 (66.5 – 30 cm)

A sharp contact occurred from facies 11 to 12, the latter being a medium grey, fine sand with a mean grain size of 198 µm, that was only moderately sorted and finely skewed. The base of the facies contained numerous BSFS 5-7 mm clasts, and occasional mudstone clasts. Organic content declined to 0.7% at 65cm. A gradational contact occurred to a dark-grey facies 13, with a mean grain size of 205 µm within a moderately well sorted, finely skewed grain population. Modern woody plant roots were present which when removed the sediment, yielded an organic content of 2%.

Foraminifera were not present in facies 13, and were infrequent in facies 14. Small but coarsely agglutinated, round and chambered *Ammobaculites sp,* occurred most commonly and was located on modern tidal flat to salt marsh pioneering elevations at Luttrells Bay in south eastern Tasmania (see Section 5.4). Light grey *Milliammina oblonga* were present with less frequent and largely broken fragments of *Milliammina fusca*. These species are cosmopolitan, mid intertidal to shallow subtidal inhabitants (Yassini and Jones, 1995; Hayward et al., 1999).

Tree pollen counts were highest in SU4, dominated by *Eucalyptus* in association with *Acacia*, *Eucryphia* and *Nothofagus cunninghamii*. *Acacia melanoxylon* dominate modern swamp forests in association with *Melaleuca* on fertile poorly drained sites on carbonate substrates in the region (Kirkpatrick, 1991). *Olearia* and *Phyllocladus* were present along with *Poa* and *G. sphaerocephalus*.

Stratigraphic unit 5, facies 15-16, (30 – 0 cm).

A 7 cm gradational contact occurred from facies 14 to a medium grey fine sand with frequent FBSS clasts of up to 1 cm. The clasts diminished in size and frequency upwards through facies 15. The absence of BSFS clasts and the presence of a slightly lighter grey coloured sediment defined facies 16. It was comprised of a friable matrix of fine silty sand, bound loosely by *Sarcocornia blackiana* root and was thick with *Melalueca ericifolia* leaf litter. The roots of both species had penetrated both facies.

Foraminifera occurred frequently in facies 15 and were dominated by *Ammobaculite* spp. with minor occurrences of *Trochammina inflata* and

Jadammina macrescens. The assemblage was identified at the very low marshin south eastern Tasmania, see Section 5.3.4. Foraminifera were not common in facies 16 and only *Trochammina inflata* were present which can occur with greater than 80% abundance in mid to upper salt marsh environments, (Hayward et al., 1999; Gehrels et al., 2004).

Intertidal shovel pits 1-7.

The series of shallow shovel pits at Sealers Springs Point revealed a thin, wedge shaped sand sheet with living, pioneering sea grasses on the surface. Underlying the sand sheet was a firm, pale brown to grey sandy silt facies. The sandy silt was imprinted with the root and basal shoot formation characteristic of the *Juncus kraussii* saltmarsh rush. The facies was connected to the eroded edge of the living saltmarsh that was inhabited by *J. kraussii* and other mid to upper marsh species, (see Section 3.4.2). The connection between the salt marsh facies underlying the modern tidal sand flat, and the salt marsh facies in T1SSPC4, was evident from the transect elevation survey and the stratigraphy exposed on the eroding modern marsh edge (Figure 6-12).

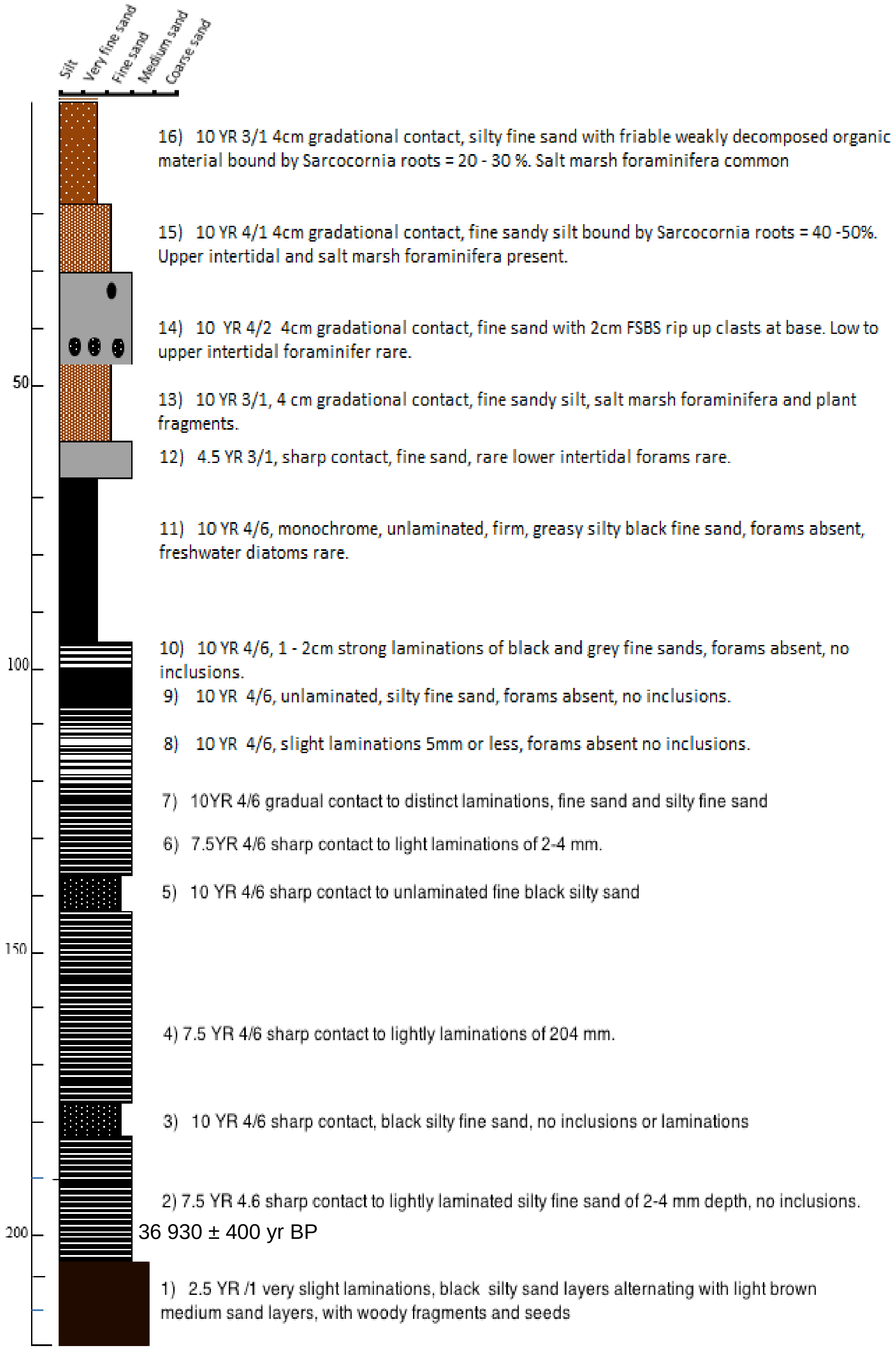


Figure 6-11 Core 4 Sealers Springs Point (T1SSPC4) log diagram with 16 facies that were differentiated by colour, texture and other lithological features such as rip up clasts. Facies 1 makes up SU1, facies 2-10 are SU2, facies 11 is SU3, 12-14 is SU4 and facies 15-16 is SU5.

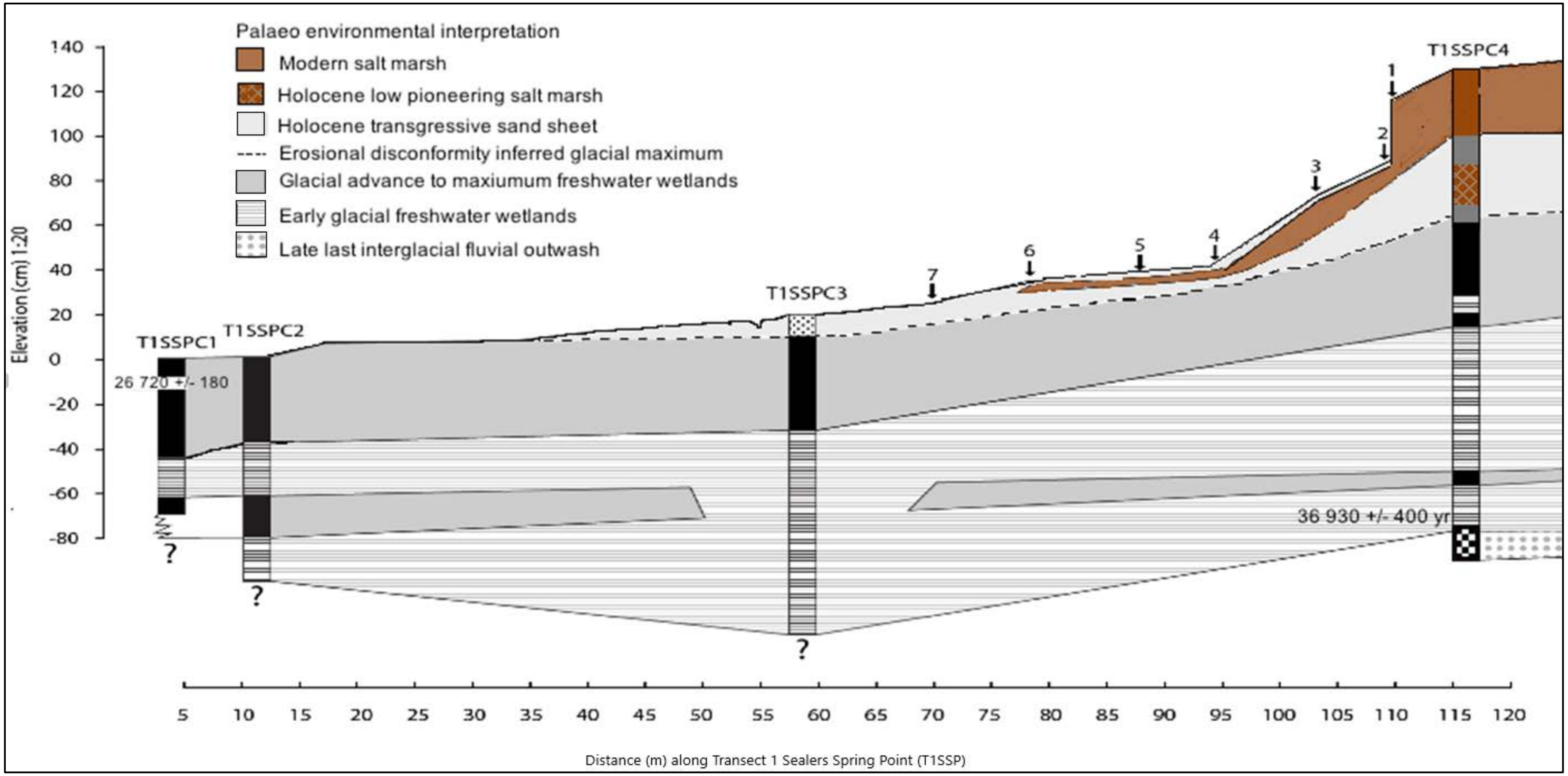


Figure 6-12 Stratigraphy of Sealers Springs Point along Transect 1 (T1SSP), showing core locations and ^{14}C sample locations. Numbers 1-7 show the location of shovel pit cores). The 7 palaeo geomorphic environments were inferred from interpretation of stratigraphic units that correlate with known climate changes over the last 40 000 years before present.

6.3.2.2 Brick Islands

Four facies were evident in the longest core retrieved from the low marsh on Transect 2 Brick Islands. They were grouped into three stratigraphic units, and their palaeo-environment interpreted in terms of its spatial connectivity with other stratigraphic units on the tidal flat and tidal channels (Section 6.4.2.4 and Figure 6-12).

Stratigraphic Unit 3, facies 1 104 – 54 cm

Facies 1 from 104 cm to 52 cm depth in core was a compacted black, silty fine sand (see Figure 6-14). The facies was devoid of marine or other biological inclusions. Mean grain size at 100 cm was 168 *u*m and was finely skewed. The particle size distribution for the Brick Island sample of the BSFS was consistent with samples from outcroppings of the deposit at the other sites and at East Montague (Figure 6-13). The transect profile with core stratigraphy shows the connection of facies 1 and its tabular form (Figure 6-15).

Stratigraphic unit 4, facies 2 54 – 23 cm

A slight gradational contact occurred over 5 cm to facies 2. It consisted of a medium grey, fine sand. Foraminifera were dominated by *M. fusca*. At the base of the facies were numerous rip-up clasts of the BSFS, ranging in size from 0.5 cm to 1.5 cm. The number of clasts declined upwards.

Stratigraphic unit 5, facies 3 – 4, 23 – 0 cm

A 4-cm gradational contact occurred from facies 2 to a brown sandy silt facies 3, that was held together by dense *Sarcocornia blackiana* root mat. It was only 15 cm deep until a small gradational contact of 3 cm occurred to a light brown fine sandy silt that was only 5 cm deep, with less root material present.

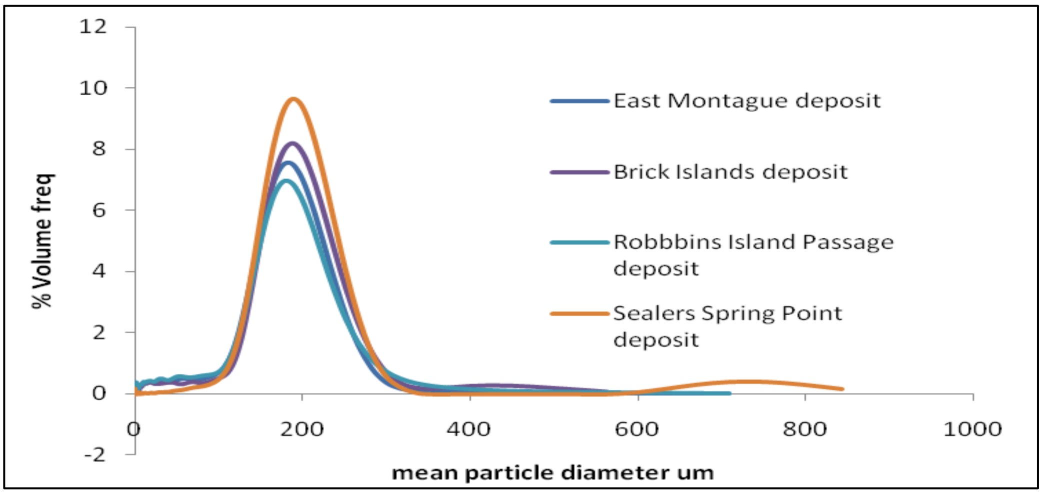


Figure 6-13 Particle size distribution by % volume (*u*m) of samples collected from outcroppings of the BSFS, showing homogenous size and distribution, although Sealers Spring Point deposit is more coarsly skewed.

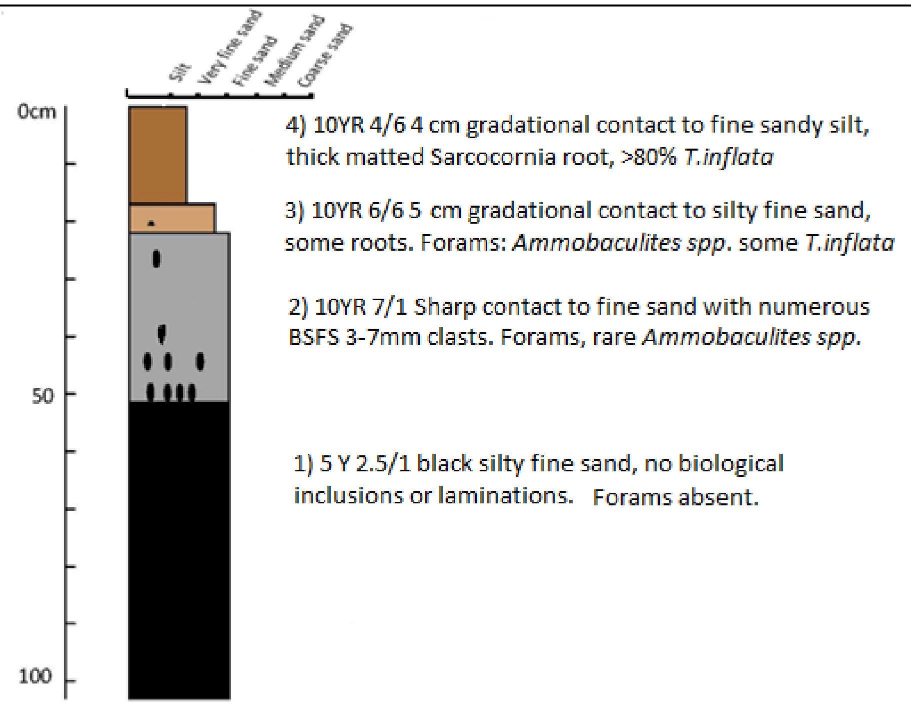


Figure 6-14 Core 2, Transect 2 at Brick Islands (C2BI) showing the numerous rip up clasts of from the terrestrial BSFS deposit, in facies 2 that contained marine microfossils.

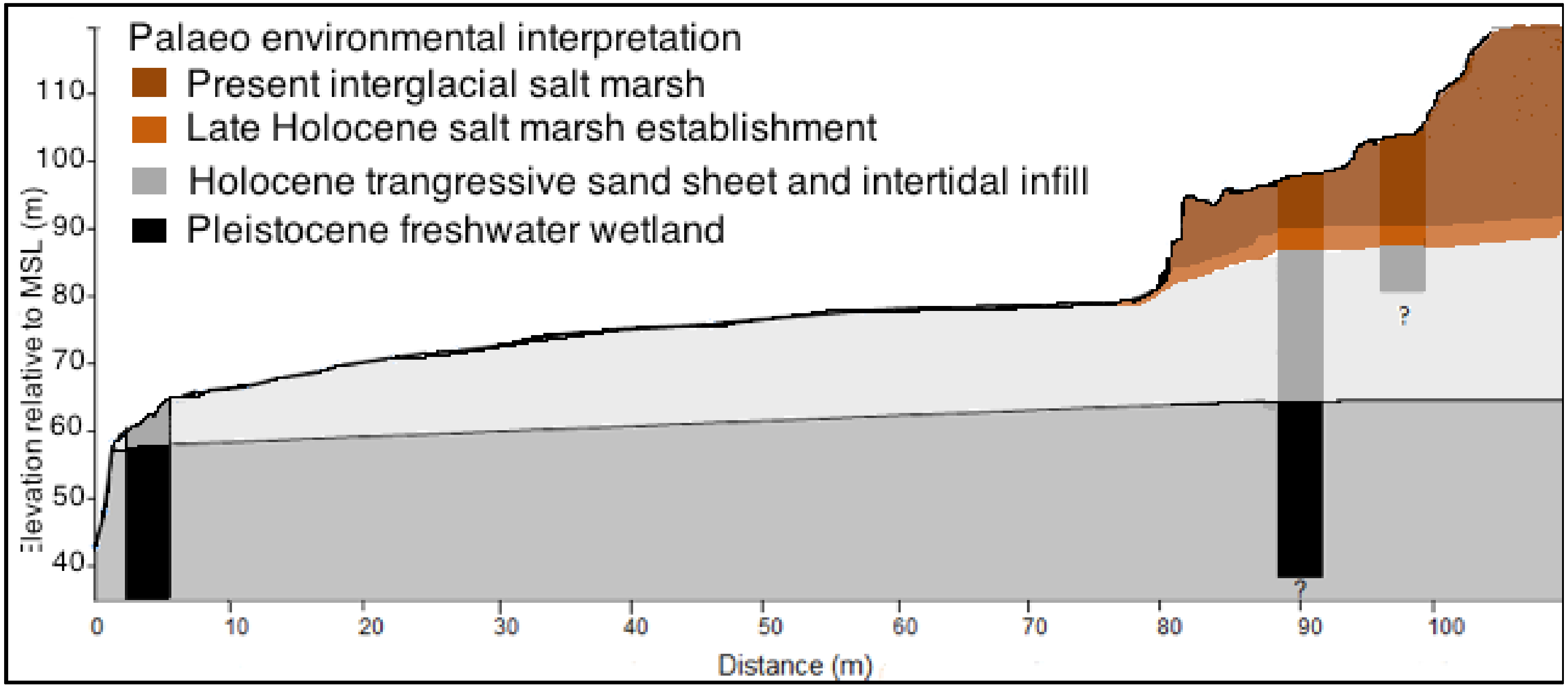


Figure 6-15 Cross section profile of Transect 2 Brick Islands showing core locations, the depth of facies 1, 2 , 3 and 4 and the location of the BSFS outcropping in the tidal creek. Interpretation of each stratigraphic unit's palaeo environment of deposition is described in Section 6.4.2.

6.3.2.3 Robbins Passage Site

The Robbins Passage stratigraphy differs from the other sites in that the BSFS was not located in the cores, or in outcroppings, and all facies contained microfossils that indicated intertidal environments of deposition. Grain size analysis showed a distinct fining upward sequence from a medium fine sand at the base of the core, to fine to very fine sand, followed by fine sandy silt in the upper facies (see Figure 6-16).

Stratigraphic unit 4, facies 1-2.

Facies 1 from 145 to 123 cm depth in core is a medium grey, medium to fine quartz sand (mean grain size 275 µm). Mudstone clasts of up to 7 mm were frequent at the base, and declined in frequency and size over 36 cm. A gradual contact was evident to facies 2 with a colour change to light grey, very fine sand (mean 105 *u*m), an absence of mudstone clasts, but inclusions of 1 - 8 mm BSFS clasts.

Foraminifera were common throughout in facies 1 which was dominated by intact *Textularia earlandi* from 139 to 92 cm depth in core. *T. earlandi* is dominant in low tidal to shallow subtidal habitats, within brackish to slightly brackish lower reaches of estuaries, and inner to middle parts of harbors and tidal inlets (Hayward et al., *1999).* Facies 2 was dominated by *Ammobaculites exigus,* the small round *Ammobaculites* sp., *Milliammina fusca* and *Milliammina oblong*a, being the same assemblage found in T1SSPC4 facies 14, SU4.

Stratigraphic Unit 5, Facies 3 and 4

Foraminifera were abundant in all samples and the upper two saltmarsh facies from 0 cm to 18 cm. At 18 cm depth in core was a sharp contact to a light brown sandy silt, and then a gradual contact at 11 cm to the fine sandy silt and clay of the saltmarsh surface. Facies 3 was dominated by the small chambered *Ammobaculites sp* and broken infrequent *T. inflata,* similar to the assemblage in T1SSPC4, facies 15, SU5. Facies 4 was dominated by *Trochammina inflata (>80%),* with infrequent *Haplophragmoides wilberti* and *Jadammina macrescens* that are known to inhabit modern mid to upper saltmarsh habitats globally (Hayard et al., 1999) and in Tasmania (Callard et al., 2011).

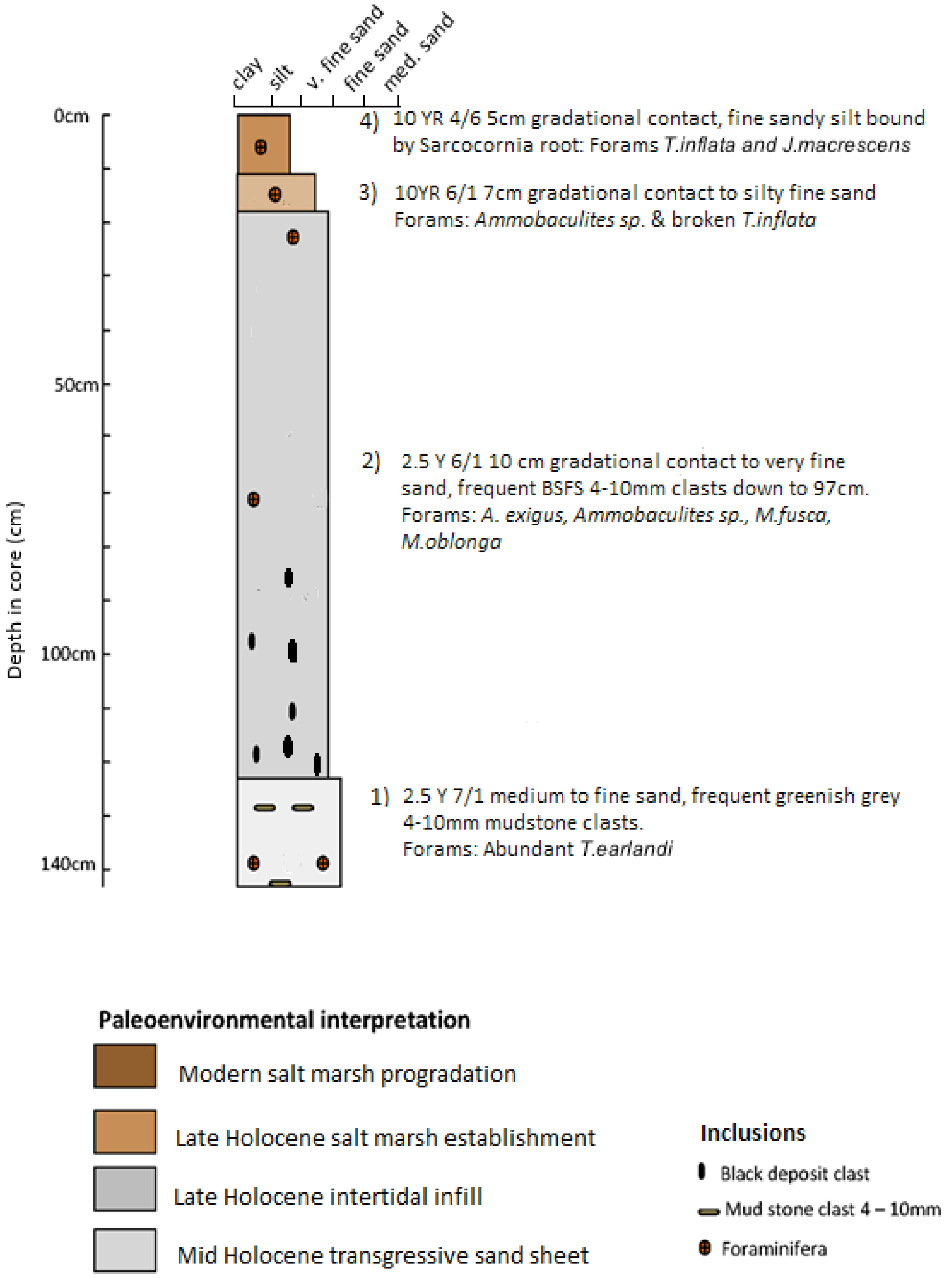


Figure 6-16 Core 1 Robbins Island Passage stratigraphy showing sediment fining upwards through the core, which is typical of an estuarine infill sequence and observable when the BSFS is not underlying at shallow depth. Facies 1 and 2 are included in SU4, 3 and 4 in SU5.

6.4 Discussion

6.4.1 Age and spatial organization of sedimentary units

The Boullanger Bay intertidal zone is made up of 3 distinct sedimentary units. All units are dominated by fine, relatively well sorted, quartzose sands, indicating a predominance of aeolian processes (Boggs, 1987). However, variable reworking processes are evident from variations in grain size distributions within stratigraphic units, from which the changing interactions between prevailing climate and sea level position over time can be inferred.

The modern salt marsh occurs from just below mean high water, to just above the highest astronomical tide mark where it transitions to *Melalueca* swamp forest. Lithological and micro fossil evidence indicates that the largest of the other two sedimentary units, the modern intertidal sand flats, are composed of intertidal infill sediment that conformably overlies a mid to late Holocene transgressive sand sheet facies. The third sedimentary unit in the intertidal zone is the black silty fine sand deposit which is either exposed, or underlies the intertidal sands. It was expected to be an early Holocene salt marsh due to its lithology and stratigraphic position, however, it yielded late Pleistocene ages that preceded the Last Glacial Maximum by about 4000 years.

Stratigraphic features in the older Pleistocene sediments are similar to sedimentary features identified in studies further inland (Colhoun, 1982; Colhoun *et al*., 1999; Colhoun et al., 2010) and elsewhere in Tasmania (Clark et al., 2011; Gehrels et al., 2012; Morrison and Ellison, 2017).

6.4.2 Chronostratigraphy and palaeoenvironments

6.4.2.1 Stratigraphic Unit 1 Late last interglacial Fluvial Outwash

Stratigraphic unit 1 is older than 36 930 ± 400 yr BP, correlating with the middle of late marine oxygen isotope stage 3, when sea levels were approximately 27 – 30 m below present MSL (Cann et al., 1988, 1993, 2000). The Tasmanian climate was cool temperate, and described as an intermediate between interglacial and glacial conditions, (Colhoun et al., 1999; Colhoun et al., 2010) with relatively high precipitation, compared to late MIS 3, (Mackintosh, 2006). At Pulbeena Swamp a wet period is indicated at around 40,000 years by a freshwater marl deposit and *Melaleuca* and *Leptospermum* wet coastal woodland communities (Colhoun, 1982).

The stratigraphic evidence in this study suggests that the environment at Sealers Springs Point was probably influenced by freshwater inputs of variable energy, indicated by an absence of any marine micro or macro fossils and the presence of seeds and plant fragments. Pollen analysis in the present study suggests proximity of the site to tussock grasslands dominated by *Poa,* buttongrass moorlands and wet coastal scrub communities. The palaeo-ecological evidence is in good agreement with the climate proxies, indicating relatively high precipitation during the transition from interglacial to glacial conditions described above (Colhoun et al., 1999; Colhoun et al., 2010).

The relative heterogeneity of the grain size, shape and lithology is characteristic of fluvial deposition. Fluvial deposits tend to be poorly sorted and coarsely skewed, due to the wide range of particle sizes available for transport by streams, and various sizes can be deposited when energy levels wane (Boggs, 1987). The inclusion of some lithic components which were rare or absent in any other facies, also indicates a fluvial outwash environment, whereby material has been transported from the land surface, rather than reworked from marine sources.

6.4.2.2 Stratigraphic Unit 2, end of the last interglacial to the early glacial, fresh water wetlands

Stratigraphic unit 2, from facies 2 through to 10 were deposited after 36 900 yr BP to before 26,720 yr BP. The 10, 000 years, span a transition period between the end of the last interglacial, to the early last glacial period. In north western Tasmania, a progressive temperature and moisture decline occurred after 35 000 yr BP, indicated by a reduction in wet forest communities and an increase in *Poaceae* and *Asteraceae*, which become competitive under increasingly dry conditions and low temperature (Mackintosh, 2006). Falling water tables on the north western coastal plain with sea level retreat would also have been occurring (Colhoun *et. al*., 1982; Colhoun and Shimeld, 2012). This is consistent with the pollen record from the present study which showed a decline in wet forest and woodland tree species, and persistence of buttongrass moorland and *Poa* grasslands.

The geomorphic processes controlled by declining temperatures and associated cold dry winds are evident by the series of laminations that indicate alternating seasonally warm periods suitable for biological activity and organic sediment accretion, and seasonally cold periods where aeolian sand deposition, sourced from the exposed coastal plain by falling sea levels, dominates. The declining seasonality upwards through the core indicated by declining laminations, is consistent with a gradual climate change that would be expected during the transition from an interglacial, to more extreme conditions of the early glacial phase.

6.4.2.3 Stratigraphic Unit 3, Glacial Advance to the Last Glacial Maximum freshwater wetlands

Stratigraphic unit 3 is comprised of only facies 11, the BSFS which is devoid of laminations. The geologic age (26,700 yr BP) near its surface, precedes the period of maximum dryness for the region, commencing around 25,000 yr BP (Colhoun et al., 1982), before the Last Glacial Maximum at 24 000 to 18 000 years BP in

Tasmania, and when sea levels were around 120 m below present levels (Petherick et al., 2013; Colhoun and Schimeld, 2012).

The BSFS deposits were extensive but not located at all sites, and varied in depth and height only minimally, indicating them to be lenticular in form. It is most likely that the lenticular morphology reflects a swamp forest that would have occupied the moist depressions in the swales between sand dunes that formed during the Last Glacial Maximum when sea levels were around 120 m below present, and have since been eroded. Such processes are similar to those that formed the environments further inland within the Smithton Basin, as indicated from freshwater marl deposits of similar age (Gill and Banks, 1956; Colhoun et al. 1982).

The lack of marine microfossils, biological inclusions, or minor lithic components and predominance of well sorted, finely skewed, fine to very fine sand, indicates that the process of deposition of the quartz grains continued by aeolian transport (Boggs, 1987). The cold, arid conditions are indicated by the low pollen counts, because pollen fossils desiccate in oxidized, dried sediment. The extent of the present Quaternary sands that drape the Smithton Basin, suggests that preservation of the BSFS form is probably due to burial by aeolian sands transported inland from the Bassian Plain as sea levels began to rise after the end of the Last Glacial Maximum around 18 000 yr BP.

6.4.2.4 Stratigraphic Unit 4, Holocene intertidal, transgressive sand sheet and sedimentary infill.

At BI and SSP, the sharp contact between the BSFS strata and the overlying facies is consistent with an erosional unconformity that indicates a break in the geologic record (Boggs, 1987). The distinct change from terrestrial to marine palaeo-environments, indicates that deposition recommenced when the transgressing sea reached elevations suitable for tidal inundation. Low intertidal and shallow subtidal foraminifera in facies 12 at SSP, and 1 at RIP indicates a Holocene transgressive sand sheet.

Transgressive processes are further indicated by clasts of the basal substrate, termed rip up clasts, referring to erosion and subsequent transport and deposition by the transgressing sea (Boggs, 1987). Concurrently, tidal currents mobilize, transport and deposit sands at shallow depths (Boggs, 1987). Such transgressive processes were also inferred from similarly positioned rip up clasts in NSW estuaries (Sloss et al., 2006 b) and at Luttrells Bay (Section 4.5.3). At Boullanger Bay, the contrasting sediment colours also enabled observation of the rip up clasts, which became sparser upwards, indicating declining tidal energy with shallowing water depths (Boggs, 1987)

Above the transgressive sand sheet, intertidal sedimentary infill processes are evident in facies 13 at SSP, 2 at RIP and 3 at BI by better sorted sediment occupied by upper to mid intertidal foraminifera (Boggs, 1987). Despite the predominance of fine sands, the sediment fining upwards from facies 1 to 2 at Robbins Island Passage was evident because the BSFS platform was absent, and therefore provided space for the deeper, coarser transgressive sand sheet. In comparison at Brick Islands, the BSFS platform occupied that space, and so only the intertidal infill facies was observed.

6.4.2.5 Stratigraphic Unit 5, present interglacial, modern salt marsh

At all sites, the gradational contact from the intertidal facies to the modern saltmarsh indicates a gradual change in the depositional environment (Boggs, 1987), with progressive intertidal infilling, until the tidal flat reaches an elevation and tidal inundation threshold, under relatively stable sea level, such that saltmarsh can establish, see Section 3.4.4.

The marshes at each site showed variable stability; RIP marshes were prograding, BI showed a balance of eroding and prograding salt marsh edges, and all edges at SSP were actively eroding. The prograding marsh edge at RIP suggests that sedimentary infill of the passage is creating lower wave energy, so that marsh can prograde seaward. The extent of the erosion indicated by extensive micro cliffing at SSP indicates that sea level rise rates are exceeding sediment supply and wave energy conditions have increased at the more open site. At SSP, 15 m seaward of the current micro-cliffed salt marsh edge is a recent mid to upper marsh deposit, buried under a thin sheet of intertidal sands. The deposit lies 39 cm below the present MSL which is 50 cm below the species present lower elevation limit (see Section 3.3.4) and is interpreted as a drowned salt marsh facies. Drowned salt marsh has not been detected in previous south eastern Tasmanian studies, and is probably evident at Boullanger Bay because the generally coarser sediment environment provides an observable contrast.

6.4.3 Intertidal landform evolution of Boullanger Bay

The five palaeo geomorphic units evident in this study show the present intertidal landform has been built by deposition of reworked fine quartzose sands over the last ~ 40 000 years before present. Seven distinct environmental changes can be inferred, resulting from a progression of climate and sea level driven processes (Figure 6-12). Changes 4 and 5 are the palaeo-environments inferred from the erosional unconformity between SU3 and SU4:

1) Interglacial fluvial outwash.

2) Freshwater moorland type environments that accreted by aeolian sand deposition, and seasonal biogenic silt production during pre-glacial times.

3) Further biogenic silt production in freshwater swamp environments, through to the later stages of glacial advance, and just prior to the LGM.

4) Burial by sand dunes built by aeolian sands transported from the Bassian Plain during the LGM to its end.

5) Aeolian erosion and inland transport of the sand dunes at the latter stages of the Last Post Glacial Marine Transgression.

6) Transgressive sand sheet deposition during the latter stages of the Last Post Glacial Marine Transgression to the mid Holocene,

7) Late Holocene intertidal infill and salt marsh establishment with relative sea level stability, and modern instability.

The net result of changing processes is the extensive intertidal flats that are uncharacteristic of a meso-tidal range. According to Eisma (1997), meso-tidal deposits only occur where there is shallow coastal water, an open coast or open connection to tides, and sediment available for building the tidal flats. Where there is an ample sediment supply and accommodation space is low, the tidal flats will reach a certain elevation relative to the tidal frame, and fill the most proximal areas first. Then, because there is still more sediment than space, landward progradation will occur (Coe,, 2003). The BSFS lenticular deposits have clearly created low accommodation space, and been fundamental to the broad intertidal landform evolution at Boullanger Bay.

The geochronological evolution of Boullanger Bay is consistent with the prevailing view that south eastern Australian Holocene coastal landforms are the result of interactions between the antecedent Pleistocene substrate that they overlie, the current boundary conditions (sediment supply and type), and relative sea level and climate forcing factors (Woodroffe, 2004; Sloss et al., 2007). Such interactions mean that large temporal spans have been captured at shallow depths from the intertidal zone in south eastern Australia (Nichol and Murray Wallace, 1992; Cann et al., 1999, 2006; Sloss et al., 2004, 2006a; 2010; Clark et al., 2011; Gehrels et al., 2012) and this is also the case in Tasmania. To accommodate the temporal span of this stratigraphic record, multiple ecological proxies have been necessary. Similarly long term environmental changes have been detected in the peatlands of Hudson Bay Lowlands in Canada (Hargan et al., 2015).

6.5 Conclusion

The meso tidal salt marshes at Boullanger Bay did not provide a stratigraphic record, suitable for precise sea level reconstructions of the mid or earlier Holocene. Indeed, there is a gap in the record between the Last Glacial Maximum and the late Holocene. Such palaeo environments are likely to be found in shallow subtidal environments and require mechanized coring equipment due to the predominance of coarse sediments.

Chapter 7 Conclusion

This thesis has developed a mid to late Holocene sea level history for Tasmania, far south eastern Australia, which was previously a gap in the Australian and global sea level data base. Sea levels rose rapidly from 7363 yr BP from around 1.75 m below present levels, to around – 0.5 m at 6000 yr BP and continued to rise very slowly with a period of stability at around 2000 yr BP, after which sea level rise commenced again, reaching around – 0.23 m 650 yr BP. To achieve this length of record from one site, several methods were used to piece together sea level history, commencing with qualitative reconstruction of mid Holocene sea levels, followed by development of a quantitative foraminifera based transfer function, to broadly indicate the nature of late Holocene sea levels. Both studies were informed by a baseline of the modern intertidal environment in Tasmania. A comparative palaeo-reconstruction in far north western Tasmania supported the south eastern study findings, and provided further evidence that Tasmania did not experience a mid Holocene sea level highstand, which was a major aim of this research.

As a starting point, vegetation zonation as an important salt marsh biogeomorphic feature linked to mean sea level, was examined across three geomorphic marsh types. It provided a baseline understanding of salt marsh form and function in Tasmania which was previously a gap, to a level consistent with knowledge in other regions that have enabled high resolution sea level research outcomes (Horton and Edwards, 2006; Engelhart and Horton, 2012). Further to providing a descriptive baseline, Chapter 3 developed a statistical model to quantify how site specific environmental conditions affect biogeomorphic responses to MSL which is important to supporting the ecological plausibility of palaeo-ecological research (Kemp et al., 2012). The findings showed that inundation period was the strongest predictor of biogeomorphic zonation across sites, but within sites, zonation was more strongly predicted by sediment moisture content and salinity, both being related to climate and geomorphic setting. Variation in inundation period within and between sites was greatest at the seaward edges, and evidence showed it to be a likely response to variable energies and sediment textures, related to geomorphic setting. The variation was lowest in the mid marsh between sites, supporting the view that species turnover from the low to mid, and mid to high marsh contains the most valuable fossil data for sea level change studies. However, the degree of variation shown to occur across sites as a consequence of variable sediment textures has implications for studies seeking to include regional modern analogue data sets, (Gehrels et al 2010; Wright et al., 2011; Kemp et al., 2012; Bartholdy et al., 2015) in that too wide a range in geomorphic conditions may create a modern analogue with too much variation, especially at the seaward edge of marsh which is an important palaeo environment that vertically constrains reconstructions.

An incidental and unexpected finding from conducting the regional scale salt marsh study is their consistently shallow depth. Their young age was confirmed by Gehrels et al., (2012) after the commencement of this thesis. This evidence and that from the stratigraphic analyses in Chapters 4, 5 and 6, points to a regional specific evolutionary pathway of saltmarsh, which was previously unknown. It is most likely that Tasmanian marshes did not develop earlier compared to older, northern hemisphere salt marshes, because estuaries did not receive large volumes of terrestrial sediment that major deglaciation causes, so tidal flat elevation did not reach an altitude suitable to *Sarcocornia* colonisation until recently. Furthermore, the relatively old geology and relatively low rainfall of the eastern Australian coastline further limited terrigenous sediment to the coast (Heap et al., 2004). Thus, sediment delivery from catchment erosion during the Holocene has been relatively slow, and estuarine infill has relied largely on marine sediment inputs. The slow rates provided an advantage to this sea level study in that the infill sequence captured the sea level signal. Having the extensive understanding of geomorphic evolution of Australian coastline from previous south eastern Australian studies, has enabled interpretation of site specific biological and sedimentological evidence obtained in this study.

The youthful nature of Tasmanian salt marshes showed that while they offered good, local constraint to MSL, a range of reconstruction methods and sea level indicators would be necessary to extend the temporal and spatial (vertical) scale of the reconstruction to the transgressive and intermediate infill sequence that dominated the sedimentary record. To this end, the site offering the longest, intact and obtainable Holocene record was located in Luttrells Bay; in the upper reaches of a low lying valley in the Little Swanport estuary. In this location, the primary thesis aim to reconstruct mid to late Holocene sea level was met in two parts:

The first part, (Chapter 4) provided a reconstruction of the mid Holocene using the low slope gradient of the palaeo flooding surface to obtain samples of incremental depth and age. Lithological and bio-stratigraphic evidence from facies overlying the flooding surface recorded the estuarine infill sequence with sea level rise. It was described using the geochronological framework for south eastern Australian estuaries developed by Sloss et al. (2006 a and b; 2010), building on the previous work of Thom and Roy, (1983) Dalrymple et al. (1992) and Heap et al. (2004). Furthermore, the south eastern Tasmanian study offers an actualistic model of estuarine infill for south eastern Australia, without the regressive features created by the fall from highstand on the Australian mainland.

The model’s palaeo geomorphic stage framework was similarly useful for describing the intertidal landform evolution of Boullanger Bay, a more open coast environment. Analysis of the comparative study area (Chapter 6), added further support for Sloss’s assertion that the topographic features of the late Pleistocene transgressive surface have influenced the nature of intertidal sediment infill resulting from sea level transgression in south eastern Australian estuaries (Sloss et al., 2006 a and b; 2010). In Tasmania’s case, the processes of very slow continuous sea level rise during the late mid, to late Holocene were inferred from comparable stratigraphic features across sites but in contrast did not record any reduction in marine conditions with back barrier development or sea level fall from highstand.

The approach is consistent with more traditional qualitative descriptions of sediment stratigraphy and dating of facies that represent particular geomorphic processes. However, the present study adapted the recommended practise of limiting precise salt marsh sample collection to the base of basal samples, after Engelhart and Horton (2011a, 2012). In the present study, in which salt marshes were young, and the record was dominated by lower tidal estuarine infill, the term ‘base of basal’ was applied to samples located at the base of the transgressive sand sheet facies, at its contact with the inferred late Pleistocene flooding surface. At Luttrells Bay, in an unambiguous environment of deposition and micro-tidal range, macrofossil samples provided good SLIPs, with accuracies of ± 0.3 m.

The new SLIPs from Luttrells Bay, in combination with existing SLIPs (Clark et al., 2011; Gehrels et al., 2012) provided a sea level curve for the mid Holocene showing that after a rapid rise of around 2 mm / year from around 8000 yr BP to 7000 yr BP, rates of SLR slowed significantly from 7000 yr BP to 6000 yr BP at a rate of approximately 0.45 mm per year. Sea levels at this time reached around 1 m below present levels, consistent with the global eustatic SLR (Lambeck et al., 2014). However, further north on the eastern Australian coastline, by 7000 yr BP, RSLs were from 1 – 2.5 m above present, and the highstand was maintained through the remaining mid Holocene. In contrast, from 6000 to 4000 yr BP, south eastern Tasmania experienced very stable, to slow rising relative sea levels that and were still around 0.4 m below present levels. This period is indicated in the sedimentary record by a thin, transgressive fine sand sheet overlying the late Pleistocene flooding surface. It was sampled for SLIPs to its shallowest feasible location, indicated by contamination of modern salt marsh root material which limited the temporal and spatial range of SLIPs to the mid Holocene. The later Holocene record was obviously held in the low intertidal infill facies which was interpreted as a geomorphological response to very slow rates of sea level rise, and relatively low sediment supply. However, because this facies could only provide opportunity for basal samples, a means of improving and quantifying their accuracy was sought.

To fill the late Holocene record gap, Chapter 5 sought to increase the temporal and vertical predictability of the existing salt marsh FBTF, developed earlier by Callard, Gehrels, Morrison and Grenfell (2011), to find suitable material for developing precise sea level index points. It contributes to the field of sea level reconstruction by testing a full intertidal foraminifera based transfer function (FBTF) in a cool temperate micro tidal setting as recommended by Woodroffe, (2009). The full FBTF yielded a relatively small RMSEP of ± 0.14 m and a low R^2 bootstrapped of 0.27, showing that the relationship between elevation and species distribution was too weak in the low intertidal zone. However, the relationship was strong around MSL, just above and below the marsh seaward edge which extended the vertical range of the FBTF by around 10%. However, the species lending the precision in this segment of the elevation gradient were not well preserved down core.

Acknowledging its limitations, the FBTF was calibrated with a core to generate a reconstruction and used to broadly indicate the nature of sea level change between 4000, to 650 yr BP. Using dated elevations from the mid Holocene reconstruction (Chapter 4), the late Holocene reconstruction indicated that sea level rose very gradually from approximately around - 0.4 m at 4000 yr BP, to around – 0.35 m by around 2000 years BP, at the same rate as the mid Holocene of around 0.1 mm / year. During this time the slowly drowning palaeo river valley evolved geomorphologically, with continuous low intertidal sediment infilling, such that salt marsh elevations were reached. The minor period of salt marsh establishment and subsequent drowning, supports that a slow rate of sea level rise was occurring although a minor period of still stand cannot be discounted. However, discerning a still stand from such a slow rate of sea level rise which approximates 0.1 mm / year assuming continuous rates from 6000 yr BP, would not be possible with the current methods. The timing of the minor sea level transgression needs confirmation, but occurring below the modern salt marsh facies indicates that it is around 2000 yr old. However, good chronological constraint of the changes in geomorphic processes related to MSL revealed in the Luttrells Bay cores, should be achieved from ample fossil macro molluscs in their life position, just below and above the salt marsh lens.

Notwithstanding the precise chronological gaps for the late Holocene this thesis has provided evidence that Tasmania's pattern of sea level change from 8000 to around 2000 yr BP is markedly different to the known pattern of sea level change for the rest of the Australian mainland, because at no point have sea levels exceeded present levels. It is consistent with the modelled glacio-hydro-isostatic adjustment for the region which is dominated by the water load term (Lambeck and Nakada, 1990). The most plausible explanation of sea level change is of continued, very slow rates of sea level rise from 6000 yr BP. Because Tasmanian sea levels are considered to closely reflect eustatic sea level change, this evidence lends support to the current debate that melt water addition from Antarctica did not cease at 6000 yr BP, but continued very slowly through the mid to late Holocene (Lambeck 1997; Nekada and Lambeck, 1989; Flemming 1998; Milne 2005). This work also provides a long term background rate to recent sea level rise studies in the region, showing that from 6000 years to around 2000 rates were less than 0.5, where the current rates are now at around 2.4 mm year (Gehrels et al., 2012).

This work establishes a basis for further precise sea level specific studies in Australia using the northern hemisphere approach of intertidal biological proxies, which now provides a data base of sea level index points for iterative glacio-isostatic adjustment modelling. Importantly, the interpretation of results in this study are integrated with a typically Australian approach of descriptively linking the morpho-stratigraphic features contained in the sedimentary record, to infer geomorphic process change related to sea levels.

7.2 Limitations to this study and recommendations for further work

Whilst this study provides good resolution SLIPs, more dating work is required, particularly for the late Holocene. Because the salt marsh record is short in Tasmania, the low intertidal infill facies described in Chapters 4 and 5, will likely be the key to reconstructing the late Holocene. Although only base of basal SLIPs are considered suitable for GIA model inputs (Engelhart et al., 2011a), this thesis has shown that intact macro fossils in the intertidal facies are likely to be in their life position, offering an error range of around ± 0.3 m. Aided by interpretation of the processes of facies deposition, and understanding of the elevation constraint offered by salt marsh deposits which cap it, confidence in the SLIPs is improved. These samples offer a good alternative to base of basal samples from locations where longer salt marsh records have been obtained. Therefore, rather than relying on salt marsh stratigraphy, studies in south eastern Australia should focus on identifying precise indicators in the transgressive and intermediate infill sequences.

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Appendix

Appendix 1

Taxonomic features used to identify species

ORDER – Foraminiferida Eichwald 1830

SUBORDER – Textulariina Delage & Hérouard 1896

FAMILY – Rzehakinidae Cushman 1933

GENUS – *Miliammina* Heron – Allan & Earland 1930

***Miliammina fusca* Brady 1870**

(Hayward et al., 1999 Pg. 81, Pl 1. 1, Figs 5-6; Yassini & Jones, 1995, Pg. 67, Figs 36 - 37)

Elongate, quinqueloculine, generally twice as long as wide (Hayward et al., 1999). Medium to large tests (>150 - >500 υm), ranging in colour from cream white to black brown with well sorted agglutinated grains building a rough walled surface (Hayward et al., 1999).

REMARKS THIS STUDY: Colours varied across sites, and reflected the relative content of quartz grains in the sediment. Relative elongation of the test also varied with site; the longest at Boullanger Bay and the shortest at Luttrells Bay.

ECOLOGICAL DISTRIBUTION: Cosmopolitan and common to abundant in New Zealand in the upper to middle parts of estuaries and tidal flats of enclosed harbours and inlets. Often the dominant species in low salt meadows and below MHT.

FAMILY – Elphidiidae Galloway 1933

GENUS – *Elphidium* de Montfort 1808

***Elphidium crispum* Linnaeus 1758 (in Hayward et al., 1999 Pg. 165, Pl. 17, Figs 9 – 10)**

DISTINGUISHING FEATURES: Medium to large circular test, with a biconvex profile and a narrow keel. 20 – 40 chambers in the final whorl that are not inflated.

ECOLOGICAL DISTRIBUTION: Found from tropical, sub-tropical Indo Pacific, Mediterranean, South West Pacific and eastern Australia. In New Zealand it occurs in exposed, sandy, fully marine, shallow subtidal environments (Hayward et al., 1999).

In this study it comprised > 80% relative abundance on the low tidal flat and subtidal sections of the middle to lower reaches of the Little Swanport Estuary. It occurred in association with other *Elphidium* and *Ammonia* species. At Boullanger bay it occurred on the low intertidal sand flats.

FAMILY – Trochamminidae Schwager 1877

GENUS – *Jadammina* Bartenstein & Brand 1938

***Jadammina macrescens* (Brady 1870) (in Hayward et al., 1999, Pg. 86, Pl. 1, figs 1 -3: Hayward and Gross, 2013 in Hayward et al., 2016)**

DISTINGUISHING FEATURES: Brown and thin walled, two inner chambers small and inward and outer chambers often collapsed producing raised sutures. Circular lipped secondary apertures on the apertural face only evident in larger terminal chambers that were not collapsed.

REMARKS: This species was difficult to identify from upper marsh samples because of its similarity to *T. inflata* which dominated the assemblages there. It has been referred to as *Trochammina* as a second form, that lacks the distinctive secondary apertures on the apertural face (Hayward et al., 1999). In this study the difficulty was presented by the smooth, seemingly thin brown walled, with highly uniform shaped chambers, that decreased in size inward, and were not always collapsed. Wright et al. (2011) refers to an intraspecific *Trochammina macrescens* series which are grouped together as *Jadammina macrescens*. Hayward et al (2016) provide images of *Trochammina inflata var. macrescens* (Cushman 1920) and this image is most like this variation at Little Swanport, higher in the tidal frame. *Trochammina inflata macrescens* is accepted as *J. macrescens* (Brady 1870).

FAMILY – Hauerinidae Schwager 1876

GENUS – *Paratrochammina* Brönnimann 1979 emend. Brönnimann & Whittaker 1988

***Paratrochammina bartrami* (Hedley, Hurdle & Burdett 1967) (in Hayward Pg. 86 & Pl. 2 Figs 1-3)**

Remarks: Very grainy and chambers and sutures only discernible by mounting individual on a gummed slide, wetting it thoroughly and observing it whilst drying as the moisture is removed from the sutures first and so the shape of the chambers are evident. Once dry, the umbilicus is only observable as a slight depression because it is filled with organic fines.

GENUS – *Trochammina* Parker & Jones 1859

***Trochammina inflata* (Montague 1808) (in Hayward pg 87, Pl. 2, Figs 6-8)**

DISTINGUISHING FEATURES: Inflated**,** test cream to gold to brown color, increasing in depth inward with smooth, well cemented walls. Large to small size, (> 500ʋm - 63ʋm) with 5 - 6 chambers in the outer whorl. Deep umbilicus

Remarks: Easily distinguished and does not vary between sites but can vary within sites with altitude. At Little Swanport it becomes slightly elongate than rounded. Alternatively *T. irregularis.*

***Textularia earlandi* (Parker 1952)** (in Hayward pg 87, Pl.2, Figs22-23)

DISTINGUISHING FEATURES: Small elongate test, mostly straight in outline. The 1st few chambers are planispiral coil but under the dissection microscope this is not clear but they increase in size and chambers form as biseral pairs of inflate chambers of around 8-10 although may be up to 15 (Hayward et al., 1999).

GENUS – *Quinqueloculina* (d'Orbigny 1826)

***Quinqueloculina seminula* Linnaeus 1758** (in Hayward, Pg. 99, Pl. 5, Figs 9-10; **Linne 1767** in Yassini & Jones, Pg. 85, Figs 198 - 199).

DISTINGUISHING FEATURES: Test ovate and variably sized from > 63 ʋm – > 212 ʋm. Test range from long (about one and a half times as long as wide, Hayward et al., 1999), to less than on time as long as wide but these are always in the larger test size range. The latter consistent with a elongate, rounded periphery description and images in Yassini and Jones (1995)

REMARKS: Hayward et al (1999) note a range forms approximating different species and that these are depth related. There was no evidence to indicate different species in these samples. Tests were fragile and often "uncoiled" or split longitudinally.

ECOLOGICAL OCCURANCE: In New Zealand, widespread and abundant in shallow exposed environments from mid shelf depths to mouths of inlets and harbours. Smaller thin walled juveniles sometimes in high numbers in marsh environments and probably transported in.

FAMILY – Hormosinidae Heackel 1894

GENUS - *Reophax* (Montfort 1808; emend Bronnimann & Whittaker 1980

Reophax nana (Rumbler 1913)

Descriptions: Comacho *et al.*, 2015, (Figs 1-4, 11-14); Stephanoudis et al., 2016 Pl. 2, 1 – 10; Comacho *et al.,* 2012 in Hessemen, 2016); Gross, O. 2004, in Hayward et al., 2016;

DISTINGUISHING FEATURES: Very small elongate and uniseral test, dark brown with loosely joined chamber appearance. Aperture on top of last chamber (Comacho 2015).

REMARKS: Classified as *R. nana* because isolated to low limit of salt marsh as described by Palaeo Electronica found in the Guadiana Estuary on the Iberian Penninsula.

Difficult to discern chambers due to dark brown colour and mixed sediment sizes creating, variably rough and irregularly shaped chambers but 3-4 may be visible, often built around silica sponge spicule although spicules rarely survive “picking”.

ECOLOGICAL DISTRIBUTION: Not commonly described in shallow benthic assemblages. Usually an accessory species limited to low salt marsh zones.

FAMILY – Haplophragmoididae Mayne 1952

GENUS – *Haplophragmoides* Cushman 1910

Haplophragmoides wilberti **Anderson 1953 (Hayward et al., 1999 Pg. 84, Pl. 1, Figs 25-26)**

DISTINGUISHING FEATURES: Small to Large test (>63 υm - > 500 υm), cream to very light brown colour, finely agglutinated and smooth walled. Broadly rounded periphery but not inflated. Inovute, planispiral with 6 – 10 chambers evident in the outer whorl and sutures can be straight to sigmoidal.

ECOLOGICAL DISTRIBUTION: Cosmopolitan in brackish environments. Common to abundant in salt marshes, high salt meadow in the upper reaches of estuaries and harbours. Lives intertidally, above MSL and most abundant above mean high water, (Hayward et al., 1999).

Haplophragmoides australensis **Albani 1978, (Fig. 6. A 2)**

DISTINGUISHING FEATURES: Small test, cream to light brown, smooth walled involute test with 5 – 6 moderately inflated chambers and depressed sutures. The test is slightly elongated at the final chamber end. The aperture is a small slit at the base of the final chamber.

GENUS – *Trochamminita* Cushman and Brönnimann; emend Saunders 1957

Trochamminita salsa **Cushman & Brönnimann 1948 (Hayward et al., 1999, Pg. 84, Pl. 1, Figs 30 - 32)**

DISTINGUISHING FEATURES: Medium size, cream to gold to brown coloured planispiral test with 6-7 inflated chambers in the final whorl followed by other chambers that form rectilinearly, but are irregular, globulose and sometimes collapsed.

REMARKS: Once mounted the inflated specimen usually collapses and the umbilicus side can resemble a figure 8 configuration.

KNOWN ECOLOGICAL DISTRIBUTION: Sporadic occurrences globally. In New Zealand it occurs abundantly in the brackish upper reaches of estuaries and at the limits of saltwater intrusion. Abundant intertidally but can occur subtidally**.**

FAMILY – Rotaliidae Ehrenberg 1839

GENUS – Ammonia Brunnich 1772

Ammonia beccari (Linnaeus, 1758)

DISTINGUISHING FEATURES

REMARKS: A. beccari was less common than A. aoteana and distinguished by its depressed sutures that made them distinct, particulary with weathered fossils. It was much larger than *A. aoteana.*

Ammonia parkinsonia f. aoteana (Finlay 1940)

DISTINGUISHING FEATURES: Low conical test, biconvex, chambers in a trochospiral arrangement, test with low conical test. Spiral side with distinct sutures that are flush or slightly raised. 8 -11 chambers.

REMARKS: Was distinguished from *A. beccari* by smaller size and not depressed sutures

FAMILY – Lituolidae de Blainville 1827

GENUS – *Ammobaculites* Cushman 1910

***Ammobaculites exigus* Cushman and Bronniman 1948 (Hayward et al., 1999, Pl. 1, Figs 19 – 20)**

DISTINGUISHING FEATURES: Small to medium sized early round coiled portion with slightly inflated chambers. Sutures not distinct until the later rectilinear portion. Generally consists of large particles ranging from black to white creating a roughly agglutinated appearance.

REMARKS: Broken rectilinear portions distinguished from *P. findens* by colour and texture where the later is paler, more finely agglutinated and aperture smaller.